

LUNDQUA



THESIS

WEICHSELIAN LITHOSTRATIGRAPHY
AND GLACIAL ENVIRONMENTS
IN THE VEN-GLUMSLÖV AREA,
SOUTHERN SWEDEN

Adrielsson, Lena

Lund 1984

Volume 16

Lund University, Departement of Quaternary Geology



LUNDQUA THESIS 16

WEICHSELIAN LITHOSTRATIGRAPHY AND GLACIAL ENVIRONMENTS IN THE VEN-GLUMSLÖV AREA, SOUTHERN SWEDEN

by

Lena Adrielsson

fil. kand., Hb

Lund 1984

Volume 16

Lund University, Departement of Quaternary Geology

**WEICHSELIAN LITHOSTRATIGRAPHY AND GLACIAL
ENVIRONMENTS IN THE VEN-GLUMSLÖV AREA,
SOUTHERN SWEDEN**

by

Lena Adrielsson

fil. kand., Hb

Avhandling

Att med tillstånd av matematisk-naturvetenskapliga fakulteten vid Lunds universitet för avläggande av filosofie doktorsexamen offentligen försvaras i Farmakologiska institutionens föreläsningssal, Sölvegatan 10, Lund, fredagen den 8 juni 1984 kl. 10.15.

Organization LUND UNIVERSITY		Document name DOCTORAL DISSERTATION	
Department of Quaternary Geology		Date of issue May 1984	
		CODEN: LUNDBDS/CNBGK-1015)1-120/(1984)	
Author(s) Lena Adrielsson		Sponsoring organization	
Title and subtitle Weichselian lithostratigraphy and glacial environments in the Ven-Glumslöv area, southern Sweden			
Abstract This thesis comprises 1. a lithostratigraphic classification of glacial sequences in the Ven-Glumslöv area and 2. a stratigraphic synthesis with a palaeo-environmental reconstruction. The following informal lithostratigraphic units were recognized: Uranienborg member, Ålabodarna member, Glumslöv member, Västernäs member, Laebrink member, Kyrkbacken till and Ögnäs sand. The stratigraphic synthesis is: Ålabodarna till member represents a glacial advance from the Kattegatt area. Deposition was by proglacial gravity flow resedimentation in a glacial lake, dammed by the glacier. After ice recession, the Glumslöv member was deposited in a glacial lake and a sand-dominated braided stream environment. Extensive glacitectonic deformation preceded the deposition of the Västernäs till member, a subaerial deglaciation sequence, deposited by an active glacier during climatic amelioration. Permafrost conditions prevailed during the next glacial advance, when Laebrink till member was deposited. The upper part of the member indicates a deglaciation during sea level transgression from 10-13 m a.s.l. to >40 m a.s.l. The deposition of Kyrkbacken till is connected to transgression maximum values in the area. The till was deposited either from a true glacier or from drifting icebergs. The glacial history is connected to the late Weichselian glaciation.			
Key words Ven island, lithostratigraphy, palaeoenvironmental reconstruction, glacial environment, glacial lake sediment, sand-dominated braided river, till, basal temperature of glacier, sea level transgression			
Classification system and/or index terms (if any)			
Supplementary bibliographical information 600 copies		Language English	
ISSN and key title 0346-8976 LUNDQUA THESIS		ISBN	
Recipient's notes	Number of pages 120		Price 35 SKr
	Security classification		

Distribution by (name and address)

Department of Quaternary Geology, Tornavägen 13, S-223 63 Lund, Sweden

I, the undersigned, being the copyright owner of the abstract of the above-mentioned dissertation, hereby grant to all reference sources permission to publish and disseminate the abstract of the above-mentioned dissertation.

Signature Lena AdrielssonDate 1984-05-21

Contents

1	Introduction	1
2	General physiographical and geological setting	2
3	Earlier concepts on the stratigraphy of the Pleistocene deposits ..	3
4	Methods	4
4.1	Field work	4
4.2	Facies states	4
4.3	Grain size analysis and grain size parameters	6
4.4	Gravel analysis and statistical treatment	6
5	Lokal stratigraphy	8
5.1	Site descriptions	8
5.1.1	Site 1	8
5.1.2	Site 2	13
5.1.3	Site 3	15
5.1.4	Site 4	17
5.1.5	Site 5	22
5.1.6	Site 6	23
5.1.7	Site 7	24
5.1.8	Site 8	26
5.1.9	Site 9	29
5.1.10	Site 10	31
5.1.11	Site 11	34
5.1.12	Site 12	34
5.1.13	Site 13	37
5.1.14	Site 14	38
5.1.15	Site 15	41
5.1.16	Site 16	43
5.1.17	Site 17	44
5.2	Lithostratigraphic correlations between the sites	47
5.2.1	Correlations based on stratigraphic position, facies states and unconformities	47
5.2.2	Comparison of directional properties in Facies E1 units	47
5.2.3	Correlations based on the gravel analysis results	49
5.2.4	The reconnaissance of reworked rock clasts	50
5.2.5	Comparison of grain size composition of Facies E1 units	52
5.2.6	The correlation of the boring log to the lithostratigraphy of sites 1 to 16	53
5.3	Lithostratigraphic classification	53
6	Stratigraphic synthesis, sedimentary processes and palaeo- environmental reconstruction	55
6.1	Ålabodarna member; gravity flow sedimentation in a proglacial lake	55
6.1.1	Provenance of material	56
6.1.2	Sedimentary characteristics and interpretations	56
6.1.3	Environmental interpretation	61
6.2	Glumslöv member; a braided river plain and a glacial lake environment	62
6.2.1	Sedimentary characteristics and interpretations	62
6.2.2	Environmental interpretation	81
6.3	Möllebäcken glaciotectionics	83
6.3.1	Previous investigations	83
6.3.2	General considerations of glaciotectionics	83
6.3.3	Deformation characteristics and interpretations	83
6.3.4	Environmental interpretation	84
6.4	Västernäs member; the deglaciation of a temperate, active glacier	85
6.4.1	Sedimentary characteristics and interpretations	85
6.4.2	Environmental interpretation	95
6.5	Laebrink member, units 24-28: an ice advance across a permafrost area	96
6.5.1	Sedimentary characteristics and interpretations	96
6.6	Laebrink member, units 29-32, and Kyrkbacken member; the disintegration of an ice during a marine transgression	104
6.6.1	Sedimentary characteristics and interpretations	104
6.6.2	Environmental interpretation, units 24-32	110
7	Regional implications	112
8	Acknowledgements	114
	References	115

1 Introduction

Stratigraphic correlations of glacial sequences have traditionally been based on a combination of till stratigraphy and iceflow-directional indications. A close study of modern stratigraphic works, dealing with the southern border zone of the Pleistocene Scandinavian ice sheets, shows that iceflow directional indications, such as striae, provenance of rock components, fabric in lodgement tills and deformation directions in glacitectonics, are the predominating correlation instruments (e.g. several of the articles in Ehlers 1983). Glacial advances and ice streams have been recognized and correlated over extensive areas, sometimes up to thousands of kilometres (see maps of ice border lines in NW Europe in e.g. T. Nilsson 1983). Ice-flow indicators are, however, in many cases produced in the ice marginal zone and cannot be taken as an indication of a regional ice flow pattern.

The traditional ice sheet models (Torell 1864, 1872, Flint 1957, 1971) have been criticized, and new conceptions of rate and mode of ice sheet growth have been presented (e.g. Andrews and Mahaffay 1976, Andrews and Miller 1979, Andrews 1982, Boulton et al. 1977, Ives et al. 1975, Lagerlund 1980). A focusation and application of theoretical glaciation and deglaciation models on till stratigraphy in southern Sweden was made by E. Lagerlund (Lagerlund 1980, Berglund and Lagerlund 1981). A new glaciation model grew out of a critical analysis of the traditional Holmström-Munthe (1904, 1920) ice-stream model (Lagerlund 1977), which ever since Å. Mattsson's (Mattsson 1960) investigations on glacial striae in Skåne and on Bornholm had become more and more unsuitable. The new model implies a growth of the Weichselian ice sheet by successive outlet surges in proglacial waterfilled depressions, alternating with the growth of marginal domes on the ice-surge masses. A further advance by surging of the marginal dome is a theoretic possibility, and the repeated pattern of marginal domes and collapses by surging let the ice margin gradually move forward (see Lagerlund 1980 for further explanation). If this model is applied to the ice flow directional pattern, the conclusion must be that an outlet surge and the growth of a marginal dome result in a complex local ice flow pattern, which varies between the different parts of an ice-dome area. Thus, the ice flow directional pattern cannot be considered a unifying pattern and cannot be used for regional correlations.

A strict lithostratigraphic classification and correlation, which follows the recommendations in the International Stratigraphic Guide (Hedberg 1976), cannot always be applied to regional correlations of glacial sequences. The classification implies the existence of strata with a well defined geometry, and the lithostratigraphic unit should be capable of being mapped in the field (Hedberg 1976). The complicated glacial environment with a variety of depositional, erosional and deforming processes, especially at the margin of a glacier, causes strata of very restricted lateral extent, and often with complicated interrelations. Repeated minor ice oscillations during ice advances or retreats may produce similar, complex sedimentary sequences over large areas. This might be wrongly interpreted as several major advances, unless the stratigraphy of the region is very well known.

The conventional lithostratigraphic correlation methods are thus, for several reasons, difficult or unsuitable to use in the regional correlations, and it is necessary to look for other unifying criteria.

The new conceptions of rate and mode of ice sheet growth also include aspects on the dynamic and thermal regime of the ice sheets and ice domes. The climatic and physio-graphical conditions, which govern the ice behaviour, are of a regional character, and it is rather the environmental conditions than the deposits themselves that are the unifying character.

The recent research on glacial sediments and especially on processes at modern glaciers has radically changed the possibility of reconstructing the physical palaeo-environment from ancient glacial sediments. The INQUA Commission on Genesis and Lithology of Quaternary Deposits (Schlüchter 1982) has drawn up a genetic classification system and has compiled criteria for recognition of glacial sediments, formed by different processes in a variety of glacial environments.

The aim of the present work has been to make a detailed lithostratigraphic study of a minor area, and to use the lithostratigraphy as a framework for a palaeo-environmental history, based on sedimentological studies and a modern till genetic classification system. Finally, the regional implications of the environmental conditions have been evaluated.

The investigated area (Fig.1) has a very complex stratigraphy, exposed in extensive cliff sections, which provides good possibilities for lithostratigraphic and sedimentological studies. The area also has a strategic position in a region with a complex ice stream pattern during the Weichselian glaciation.



Fig. 1. Location map, the arrow indicates the field-study area. The heavy line shows the maximum extent of the Weichselian ice-sheet in the south-western Baltic area.

2 General physiographical and geological setting

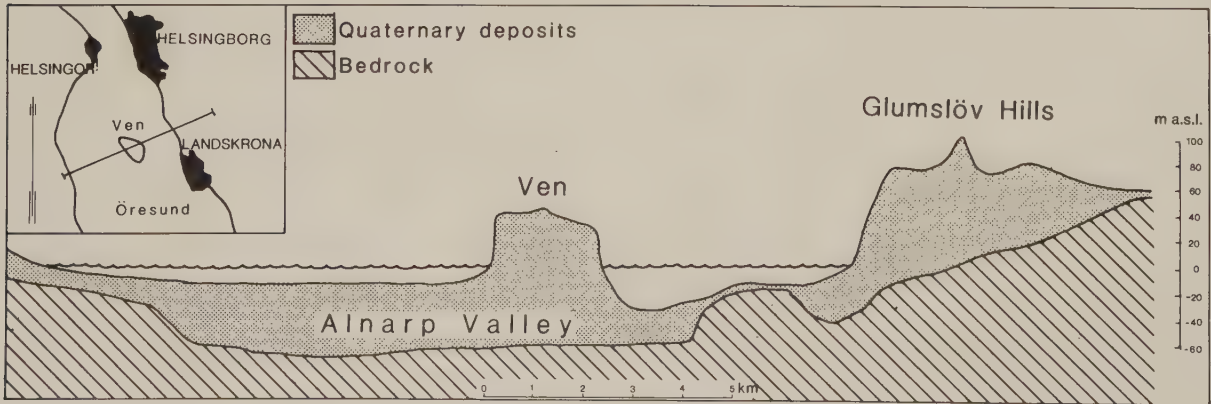
The bedrock surface lies at about -60 m (below sea level) at Ven and elevates to +50 m (above sea level) about 10 km towards the north-east. This rather steeply inclining bedrock surface is a combination of two different tectonic features.

Ven is situated in the middle of the Alnarp Valley (Fig. 2), a graben (Sorgenfrei 1945) trending north-west from the Baltic Sea south of Skåne to the Kattegatt Sea north of Sjaelland (Holst 1911, V. Milthers 1922). The valley bottom is almost plane at a level of -60 m to -65m and is about 10 km wide in the Ven area. The continued bedrock elevation on the mainland is part of the Thornquist Line (Fig. 2) (Thornquist 1908), which reaches Skåne approximately at Helsingborg and continues towards the south-east into the Romeleåsen uplift. This fault line separates the Danish Embayment from the Fennoscandian Shield. On the south-west side of the line, the bedrock lithology is rather uniform and consists of Cretaceous and Tertiary sedimentary rocks. The Fennoscandian Border Zone,

on the other side of the line, is a bedrock mosaic of uplifts and troughs with rocks ranging in age from Archean to Cretaceous (Bergström et al. 1973).

The Quaternary deposits have a considerable thickness in the area, and up to 100 m of unconsolidated sediments have been recorded from borings (Adriellsson et al. 1981) (Fig. 3). The morphology of the Alnarp Valley cannot be traced in the present surface relief, and the strait of Öresund crosses the buried valley obliquely.

The island of Ven rises above the strait as a flat plateau with steep, sometimes almost vertical cliff shores which in some parts are more than 35 m high. A conspicuous feature of morphology on the mainland is a row of hills which trends north-west from the coast across Glumslöv and Rönneberga. The hills lie on the bedrock slope and are built up of Quaternary deposits. The coastline cuts the western part of the hills, and the cliffs exhibit the internal construction.



3 Earlier concepts on the stratigraphy of the Pleistocene deposits

The dislocated strata in the cliffs were observed already by Oerstedt (1844), and ever since his first brief description they have been of interest to earth scientists.

The multiple till beds and the glaciectonic structures, with evidence of different ice flow directions, constituted a brick in the construction of the early model of the Scandinavian Ice Sheet (Torell in Holmström 1865). Two distinctly different ice flow directions were registered on the basis of the petrography of clasts in the tills (Holmström 1865, 1874, Torell 1872, Erdmann 1873, 1874, 1881, 1883). The older ice stream came from N to NE and the younger from S. This pattern was later correlated to an ice-stream model for South Scandinavia, set up by Holmström (1904). The model comprises: (1) The Old Baltic Ice with an ice flow from SE-E, (2) The NE Ice with an ice flow from N, changing into NE during the main phase and locally into SE during the deglaciation, and finally (3) The Low Baltic Ice, a readvance from S in south-western Skåne. The two ice flow directions, interpreted from the sediments in the Ven-Glumslöv area, were correlated to the NE Ice and the Low Baltic Ice, and investigators during the 20th century made the same correlations (Holmström 1912, Madsen 1917, Hansen 1933, 1940, Wennberg 1949, Johnsson 1956, 1958, 1962, 1963, Markgren 1961, Rasmussen 1973a, 1973b, 1974).

The traditional ice-stream model (Holmström 1904) and the commonly accepted Fennoscandian ice-sheet model (Torell 1864, 1872) have been seriously criticized and shown

to be unreliable and not in accordance with modern concepts on ice sheet dynamics (Lagerlund 1977, 1983, Berglund and Lagerlund 1981).

A special attention has been paid to the extensive strata of silty clay and the fine sand, which are exposed in the cliffs and the clay pits. The clay contains scattered shells of marine molluscs, some of them indicating temperate climatic conditions and some indicating arctic conditions. Some parts of the silty clay also contain dispersed clasts, and the combination of marine shells and large clasts has led to the most contradicting genetic interpretations as marine preglacial, interglacial or interstadial sediments (Munthe 1896, Holst 1911, Holmström 1912, Ödum 1933), pro-glacial lake sediments (Hansen 1933, 1940, Johnsson 1956, 1958, 1962, 1963), sub-glacial lake sediments (Wennberg 1949), glacio-marine drop-till (Rasmussen 1973a, 1973b, 1974) or till (Torell 1872, Erdmann 1873, 1874, 1881, 1883).

The contradicting genetic interpretations have also resulted in a variety of chronostratigraphic correlations, and the silty clay and the sand have been given ages of preglacial (Holst 1911), Eem interglacial (Ödum 1933), Weichselian interstadial (Mörner 1969, Nilsson 1973) and Weichselian stadial (Wennberg 1949, Johnsson 1956 etc, Rasmussen 1973 etc.). Mörner, in his chronological subdivision of the Weichselian ice age, even gave the interstadial at approximately 30,000 to 40,000 years B.P. the name: the Glumslöv Interstadial (Mörner 1969 p. 38).

4 Methods

The investigation was performed in two steps. The first step concerns the lithostratigraphy of the Quaternary sediments in the area and resulted in a lithostratigraphic classification which is based on the recommendations in the International Stratigraphic Guide (Hedberg 1976). The second step comprises the interpretation of the sedimentary processes and the palaeoenvironmental reconstruction. The two steps are dealt with separately, mainly in order to emphasize that the lithostratigraphical investigations have been the basis and a prerequisite for the palaeoenvironmental reconstruction.

4.1 Field work

The field work began with a reconnaissance of outcrops along 12 kilometres of coastal cliffs and in the former clay pits on the island of Ven and along the mainland coast. The sediments in the cliffs are mainly covered by slumps and vegetation, and the investigations were concentrated to 13 cliff sections and 3 clay pits (Fig. 5). Photographs of the cliff sections were taken from a small fishing-boat, and the enlarged photographs were put together in strips. Lithological units were recognized on the basis of facies states, and boundaries of individual units were traced in the field and drawn on the photo-strips. Lithological units were recognized on the basis of facies states, and boundaries of individual units were traced in the field and drawn on the photo-strips. Major faults and folds were also drawn.

Samples for grain size analyses and petrographical analyses were collected from the sediments, and fabric analyses were performed at different levels in the till units, usually in connection to the sampling points. The orientation of elongated pebbles was measured, and 50 to 100 readings were made in each set. The results are plotted as rose diagrams at the section charts. Fabric analyses, used in the genetic interpretation of the sediments, comprise measures of orientation and dip of longaxes. Prolate-shaped clasts were preferred to blade-shaped clasts, but the restricted supply of pebbles in some of the diamictons did not permit a discrimination of blade-shaped clasts. 25 to 50 readings were made in each set, and these analyses are displayed in stereographic projections in the second part of this work.

Other directional indicators were also investigated. Striae directions were measured on boulders within the individual till units and in the contact between the different units. Glacitectonic features, described by Rasmussen (1973a), were checked and supplemented by new observations. The basic material for the lithostratigraphical interpretation was collected during this first stage in the field work.

The second part of the field work comprised detailed studies of representative sequences from the different stratigraphic units. Logs of sedimentary successions were recorded. Primary sedimentary structures and deformation structures were studied in detail, and the studies were combined with a closely spaced sampling for grain size analyses. Close-up photographs were taken, and draw-

ings of special features were made. Data of palaeocurrent directions were collected from i.a. cross-bedded and cross-laminated sets.

4.2 Facies states

A genetic classification is consciously avoided in the lithostratigraphical descriptions. Descriptive nomenclatures for sorted sediments, based on distinguishing characteristics of texture and primary sedimentary structures, have frequently been used during the last decade. Most of the classifications are based on the main facies, stated by Allen (1970a). An adequate descriptive nomenclature for diamicton sediments did not exist until recently (Eyles et al. 1983). I have used a classification which was worked out from my own field experiences and which is based on texture, sedimentary structures, as fissility and stratification, and the type of associating sediments in the diamicton units.

Gravel Facies (Facies A). This facies represents a heterogeneous group of sediments and comprises both more extensive gravel beds and thin beds of granules and fine pebbles, often containing intraclasts of silt and clay, in troughs or on laterally extensive scoured surfaces. Residuals of poorly sorted bouldery gravel are also included in this facies.

Sand Facies (Facies B). A subdivision into cross-bedded sand (Facies B1), planar laminated sand (Facies B2) and cross-laminated sand (Facies B3) is made in the detailed descriptions. Facies B1 consists of tabular sets (B1 tab) or trough-shaped sets (B1 tr). The terminology of laminated sand with parallel and almost horizontal laminae is confusing. Parallel laminated sand, flat-bedded sand, horizontal laminated sand, evenly laminated sand and planar laminated sand occur as synonyms in the literature. I have used the term planar lamination, as, from a linguistic point of view, it is closely related to the plane bedform which produces the lamination type.

Facies B3 comprises different types of ripple cross lamination. Type A and type B ripples, without and with preservation of stoss sides, (Jopling and Walker, 1968) are generally distinguished. Type S (Jopling and Walker 1968) with lee side and stoss side, equal in thickness, is noticed in exceptional cases and then in connection with type B ripples. The distinction between sinuoidale ripples and draped lamination (Gustavson et al. 1975) is, however, unclear, and according to Gustavson et al. (1975) the transition is gradual.

Sharp-crested symmetrical or almost symmetrical ripples are present in several sections. These ripples show a great variation, and the different structures are generally described in the text. The sharp-crested ripples are generally developed in very fine sand and coarse silt.

Alternating Bed Facies (Facies C). This facies consists of a vertical alternation of sand, silt and clay layers. The sand is found in extensive layers with a thickness equal

to or larger than the silt and clay layers. Lenticular or wavy bedding or parallel laminated units with a sand content less than approximately 50 % is referred to the silt and clay facies.

Silt and Clay Facies (Facies D). A large variation of sedimentary structures is present in this facies, such as parallel lamination cross lamination, convoluted lamination, graded bedding and homogeneous beds. The silt and clay may contain scattered clasts and intra-clasts of clay. If the ratio of clasts and intra-clasts is high, the sediment is referred to the diamicton facies.

A parallel-laminated silt and clay facies (Facies Dlam) has been stated more precisely in some of the descriptions in the second part of this work.

Diamicton Facies (Facies E) has been subdivided into a homogeneous facies (Facies E1), a crudely stratified, banded or smudged facies (Facies E2) and a distinctly stratified facies (Facies E3).

The homogeneous facies is not considered absolutely homogeneous, but smudges or lenses of divergent texture may not exceed 5 to 10 %. If the divergent material occurs as consistent partings, this is noticed by the addition of a p, which stands for parting (Facies E1p). The homogeneous facies often shows a fissile structure pattern which is noticed by the addition of an f (Facies E1f).

Facies E3 consists of layers of different texture and generally with distinct and sharp boundaries between the individual layers or laminae. The diamicton beds are often, but not necessarily, texturally different both within and between individual beds. Associating sediments may be dominated by sand with gravel and silt as minor components (Facies E3B) or by silt and clay (Facies E3D). Primary sedimentary structures are often present in the intrabeds. Facies E2 is a heterogeneous group of sedi-

ments, and is most easily expressed as all diamictons which cannot be referred to Facies E1 or Facies E3. Facies E2 may consist of only diamictic material with an obvious stratification or banding, or it may contain streaks or bands of texturally divergent material. Smudges or lenses of divergent texture may also occur. Facies E2 may be fissile (Facies E2f).

The diamictons in the area generally have a high total silt and clay content, which rarely falls below 50% of the matrix. The content of boulders and cobbles is very low. Some sediments, referred to the diamicton facies consist only of silt and clay, according to grain size analyses. A high ratio of the clay content is, however, present in the sediments as intraclasts.

The average diamicton in the area has a total silt and clay content of 50% or slightly more. Comments on the textural composition of the diamicton units in chapter 5.1 are only made if the composition diverges from the approximate average.

Table 1 shows the facies states, used in the text, and Fig. 4 shows the ornaments which have been used in the lithostratigraphic diagrams in chapter 5.1, and corresponds to the facies states.

Table 1. Facies states, Facies A, B, C and D are after Allen (1970a).

Facies A	gravel facies
Facies B	sand facies
Facies B1tr	-trough cross-bedding
Facies B1tab	-tabular cross-bedding
Facies B2	-planar lamination
Facies B3	-ripple cross-lamination
Facies C	alternating bed
Facies D	silt and clay facies
Facies E	diamicton facies
Facies E1	-homogeneous facies
Facies E1f	-homogeneous facies with fissile structures
Facies E1p	-homogeneous facies with sand partings
Facies E2	-crudely stratified, banded or smudged facies
Facies E2f	-crudely stratified, banded or smudged facies with fissile structures
Facies E3B	-distinctly stratified facies with sand as dominating, associating sediments
Facies E3D	-distinctly stratified facies with silt and clay as associating sediments

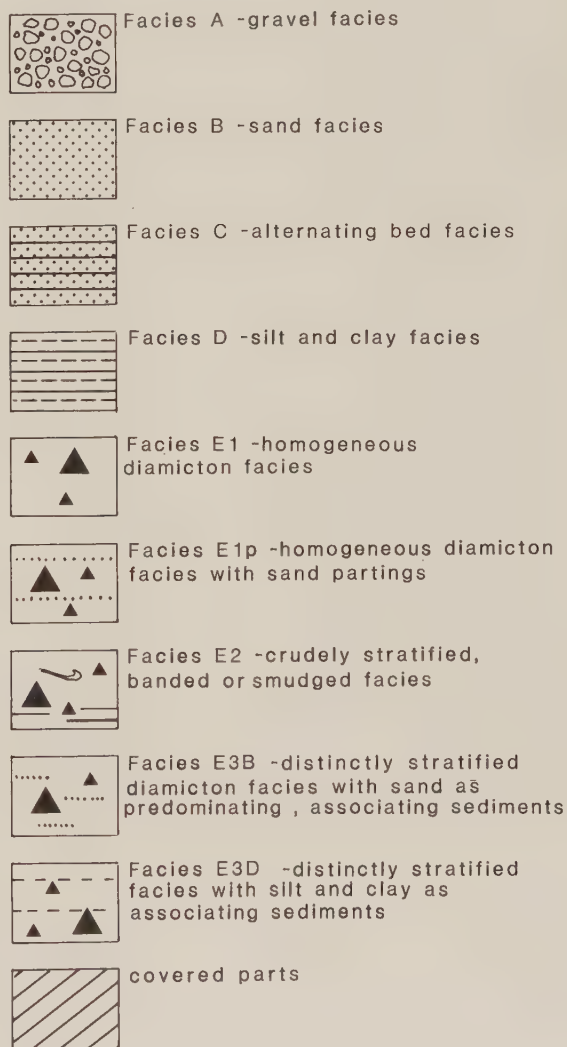


Fig. 4. Key to ornaments in the lithostratigraphic diagrams.

4.3 Grain size analysis and grain size parameters

The Ø grade scale has been used in the text. As the metric scale is most frequently used by Swedish geologists, I have converted the values to metric equivalent in Table 2. Sand and gravel fractions, 4Ø to -4Ø, were analysed by wet + dry sieving of samples of 0.5 to 5 kg weight. Samples of sorted sand were sieved with sieve interval of 1/2Ø, and diamicton samples and poorly sorted gravels were sieved with whole Ø intervals. Grain sizes finer than 4Ø (0.063 mm) were analysed by using the hydrometer method (Gandahl 1953).

Mean, standard deviation (sorting) and skewness were calculated from the cumulative curves. The statistical parameters, recommended by Folk and Ward (1957), are most frequently used in literature on sedimentology. These parameters are, however, mainly adjusted to arenaceous sediments, and start from the 5, 16, 50, 84 and 95 % values on the cumulative grain size curves. The 95 % value eliminates a great part of my samples, including almost all of the diamicton samples, as the clay content generally is too high and the cumulative curves stop before the 95 % value. To get a possibility to compare the numerous sand samples with samples from other investigations, which is important for the environmental interpretation I have used the Folk and Ward parameters (Table 2) for the examples from the extensive sequences of sorted sand sediments.

Standard deviation and skewness from Inman (Table 2) were used for the statistical treatment of samples from the diamictons and sediments, closely related to the diamictons.

Table 2. Metric and Ø grain size scales.

Millimetres	Phi (Ø) Scale	Wentworth Scale	Field classification
256	-8	Boulder	
64	-6	Cobble	
		Pebble	
4	-2	Granule	GRAVEL
2.00	-1.0	Very Coarse Sand	SAND
1.00	0.0	Coarse Sand	
0.50	1.0	Medium Sand	
0.25	2.0	Fine Sand	
0.125	3.0	Very Fine Sand	
0.0625	4.0	Coarse Silt	
0.031	5.0	Medium Silt	SILT
0.0156	6.0	Fine Silt	
0.0078	7.0	Very Fine Silt	
0.0039	8.0	Very Fine Silt	CLAY
0.0020	9.0	Clay	

4.4 Gravel analysis and statistical treatment

The petrographical composition was determined in the gravel fraction by using J. Lundquist's method (1952). The high content of poorly consolidated sedimentary rocks in many of the samples made it necessary to avoid a mechanical treatment, and the samples were washed and sieved in hot water. The gravel was divided into three grain sizes, 3-4 mm, 4-5.6 mm and 5.6 mm. The number of grains in the coarsest fraction seldom exceeded 100, which is not enough for a statistical treatment. 350 to 1500 grains were examined in the finer fractions. Only the 3-4 mm grain size has been used in the diagrams.

The grains were examined under a binocular microscope (X40). The samples contain several rock types, and the classification of the small grains was sometimes difficult. The coarsest fraction was therefore important as reference material, as the rock types were more easily recognized in this fraction. The following rocktypes were distinguished in the analyses: Crystalline rocks were subdivided into acid rocks, intermediate, basic and ultrabasic rocks. Amphibolites with garnets, diabase and basalts were separated in individual groups. Palaeozoic rocks were divided into sandstone, shales and limestones. Cambrian quartzite and quartzite sandstone were separated in the sandstone group. Shales were subdivided into black shales, grey and greenish grey shales and marly shales. Palaeozoic limestones were divided into grey and red limestones.

The lithology of the Mesozoic rocks is very heterogeneous. Different lithologies from the Kågeröd Formation were related to one group, and the Raethic-Liassic rocks were divided into sandstones, siltstones and clay, clay ironstones and coal.

Cretaceous and Danian rocks were divided into limestones, chalks and flints.

A characteristic red or violet sandstone is presented in several samples. A poorly lithified white sandstone also occurred as a cha-

Table 3. Size parametres after Folk and Ward (1957) and Inman (1952).

Folk and Ward (1957)

Mean $M_z = \frac{\phi_{16} + \phi_{50} + \phi_{84}}{3}$

Standard deviation $\sigma_I = \frac{\phi_{84} - \phi_{16}}{4} + \frac{\phi_{95} - \phi_5}{6.6}$

Skewness $Sk_I = \frac{\phi_{84} + \phi_{16} - 2\phi_{50}}{2(\phi_{84} - \phi_{16})} + \frac{\phi_{95} + \phi_5 - 2\phi_{50}}{2(\phi_{95} - \phi_5)}$

Inman (1952)

Standard deviation $\sigma_\phi = \frac{\phi_{84} - \phi_{16}}{2}$

Skewness $\alpha_\phi = \frac{\phi_{84} + \phi_{16} - 2\phi_{50}}{\phi_{84} - \phi_{16}}$

racteristic component in some of the samples. The two sandstones were also distinguished. In all, 20 to 25 rock types were distinguished in each sample.

The treatment of the results from gravel analyses has not yet been standardized, and several different types of diagrams and tables have been used in the south-western part of the North European glaciation area (Nilsson 1971, Lagerlund 1971, 1977b, Petersen and Konradi 1974, Ehlers 1979, Houmark-Nielsen 1980). A grouping of rock types has generally been made, and the amalgamation has been based on petrography, e.g. igneous and metamorphic rocks, sandstones, limestones etc. (Petersen and Konradi 1974), or on the age of rocks, e.g. Archean rocks, Cambro-Silurian rocks etc. (Nilsson 1971), or on their transport direction, e.g. from the north by Fennoscandian ice, from the south by South German rivers, from the east by North German rivers (Zandstra 1978). In his classification of associating groups, associations and types, Zandstra also paid attention to the mixing of rock types of different provenance.

The rock types, represented in the samples from the Ven-Glumslöv area, were grouped in five main groups on the basis of their petrography and age. Information on bedrock provenance is given for the individual groups or group components (see also Fig. 65).

Group A (Crystalline rocks). All types of igneous and metamorphic rocks are represented in this group. Crystalline bedrock has a great extent to the north-east in the Fennoscandian Shield, but the Archean horsts in the Border Zone, both in Skåne and in Bornholm, may also have delivered material.

Group B (Raethic-Liassic rocks). The Raethic-Liassic has its main distribution in the north-west of Skåne. Minor distributions are also found in the northern and marginal parts of the Vomb basin in the south central part of Skåne and in Bornholm.

Group C (Palaeozoic rocks). A subdivision into sandstones, shales and limestones has been made in the diagrams. The Palaeozoicum in Skåne is mainly represented by shales, and the most extensive outcrops are found in a broad diagonal band from the central to the south-east of Skåne. In most samples the Palaeozoic group is dominated by a grey limestone. Small outcrops of limestone occur in the eastern part of Skåne, but more extensive areas are found in the islands of Öland and Gotland and on the surrounding floor of the Baltic Sea.

Group D (Cretaceous and Danian rocks). A subdivision into limestones, chalks and flints is made. The Cretaceous and Danian bedrock has its main distribution in the Danish Embayment to the south-west of the Thornquist Line. Ven is situated to the west of this

line with a narrow band of Upper Cretaceous and Danian rocks to the east. Outcrops outside the Danish Embayment are found in the Bornholm, Ystad-Vomb, Båstad and Kristianstad areas (Bergström et al. 1973).

Group E (Other rocks). This group consists of the rock types that cannot be related to any of the groups described above. It also consists of undefined rocks. The content of red and white sandstones is shown in the diagrams. The provenance of the red to violet sandstone has been discussed by Lagerlund (1980). Three sandstones of different age and different bedrock distribution are possible, the Nexö-sandstone from the Bornholm area, Jotnian sandstone from the Gävle area and Devonian sandstone from the Baltic Sea to the south-east of Gotland. The white sandstone, recognized in some of the samples, probably belongs to either the Kågeröd beds or the Raethic-Liassic bedrock but is also related to group E.

The statistical treatment of the multivariate data from the gravel analyses was made by means of a numerical program, the FORT-RAN IV program PCARMODE, written by Dr. John Birks and modified for the computer at the Datacentre in Lund. The program was originally established for Quaternary pollen stratigraphic data (Adam 1974, Birks 1974, Pennington and Sackin 1975), and the method has been applied to pollen stratigraphic investigations in Lund (e.g. Birks and Berglund 1979, Björck 1979). It is, however, appropriate to other kinds of stratigraphic data, consisting of several variables. The method is based on the principal component analysis, PCA, which shows the variance-covariance structure of the data, given to the computer, and clarifies the patterns that are present in the data. For the theoretical structure and operational methodology see Morrison (1967), Davis (1973) and Jöreskog et al. (1976).

Eleven petrographical variables were used: Group A (1), group B (2), group C with a subdivision into sandstone (3), shale (4) and limestone (5), group D with a subdivision into limestone (6), chalk (7) and flint (8), and group E with a subdivision into white sandstone (9), violet sandstone (10) and others (11). The percentage share of these variables in 153 samples was fed to the computer.

The main purpose of the numerical analysis was to supplement and verify the lithostratigraphic correlation of units, which was based on physical tracing, stratigraphic position and comparison of facies states. The grouping of plots, representing samples by their eigenvectors and eigenvalues, was compared to the first lithostratigraphic correlation. Further information was obtained from the loadings of variables of component 1, which give the mutual relations between the petrographical variables in the matrix.

5 Local stratigraphy

5.1 Site descriptions

The site description will give a brief outline of the distribution of strata in the cliffs and clay-pits on the island of Ven and the nearby mainland of Skåne. A description of strata from a boring on Ven is also given. The geographical position of the investigated sections is shown in Fig. 5.

The lateral and vertical distribution of main facies states is shown in section charts. The position of features, mentioned in the text, is expressed by the number of the section and the distance in meters from the beginning of the section.

A lithostratigraphic record is given for each section or part of a section. The strata are briefly described, unit by unit, from the base to the top. The thickness of each unit is approximate and based on the average thickness in the section. The total thickness of the strata generally exceeds the height of the cliff, as the lower units are affected by glacitectonism and have been angularly adjusted. As the tectonics have already been described by Rasmussen(1973a), only supplementary measurements and features, important for the interpretation, are mentioned.

The stratigraphy and the gravel analysis results are compiled in diagrams.

5.1.1 Site 1

The site comprises the first cliff-cuttings south of the Kyrkbacken village on the west coast of Ven (Fig.5). The cliff trends S40°E, and the maximum height is 38m. The observations were made in a 620m long section(Fig.6). The part between 1:310 and 1:370 was destroyed by a landslide in the winter of 1975, and the section chart shows the cliff before the landslide. The new cutting of the landslide scar is an almost vertical wall in the upper part of the cliff. Some observations from this outcrop are described.

The lithostratigraphy of the site is dis-

played in two parts, Figs 7a. and 7b, one for the north-west portion of the cliff, 1:0 to 1:310, and one for the south-east portion, 1:310 to 1:620.

Lithostratigraphic units, section 1:0-310

Unit 1a:1(Facies E2f). The 1:0-110 part is mainly covered by slumps and vegetation and the unit has only been studied in small and scattered outcrops. The base of the unit is situated below the sea level. The fissile diamicton has a high silt and clay content of approximately 75 %. Rock clasts are rare, but occur scattered in the exposures. The diamicton contains smudges of laminated silt and intraclasts of clay.

Unit 1a:2(Facies E3D). The stratified diamicton consists of diamicton beds with a high silt and clay content, interlayered by laminated silt and clay. The diamicton beds are of two types, a silty clay matrix with intraclasts of clay and a silty clay matrix with scattered rock clasts. The two diamicton units, unit 1a:1 and 1a:2, have a total thickness of approximately 20 m in the cliff.



Fig. 5. The field-study area. The topography is given by 10 m contourlines. The position of the investigated sites is shown by their numbers. Local names, used in the text and in the lithostratigraphic classification, are shown.

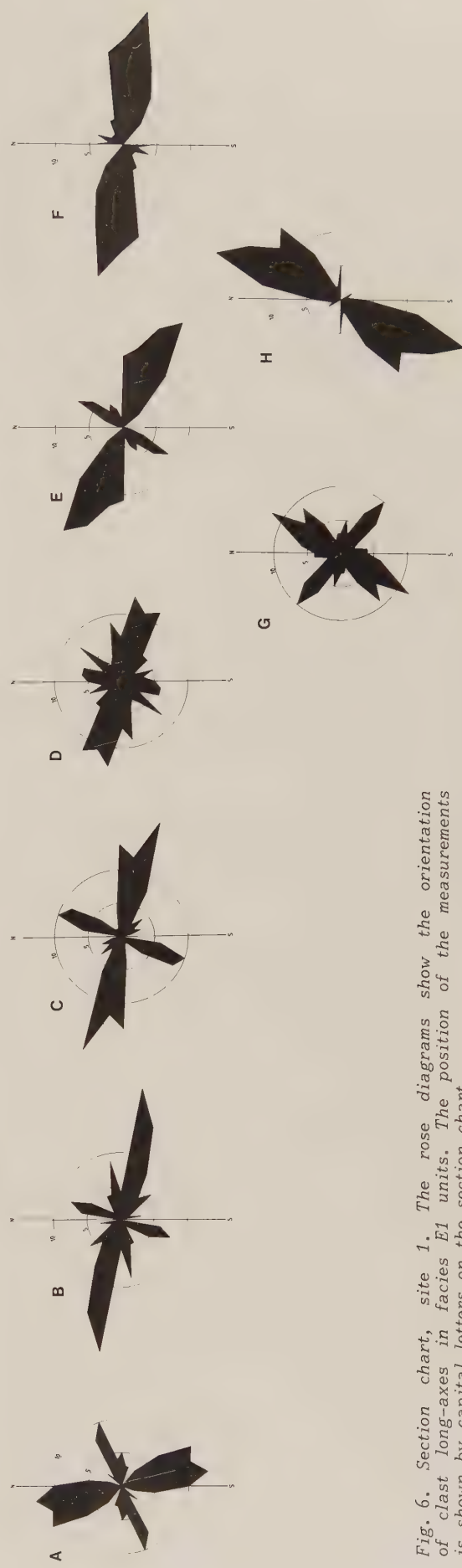
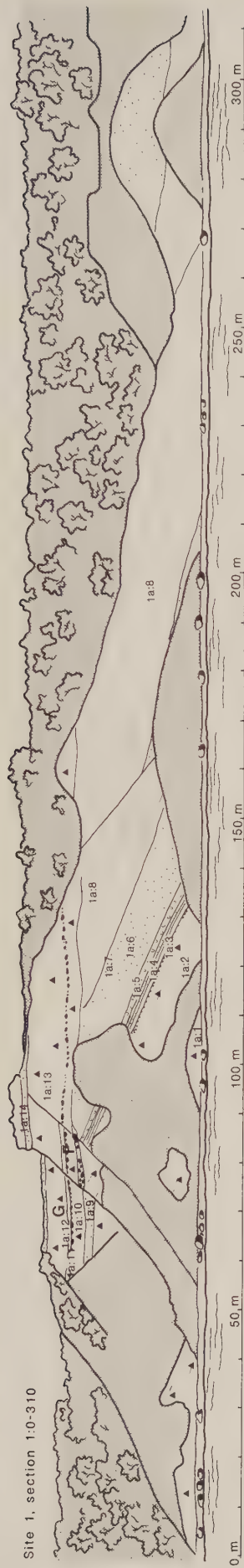


Fig. 6. Section chart, site 1. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

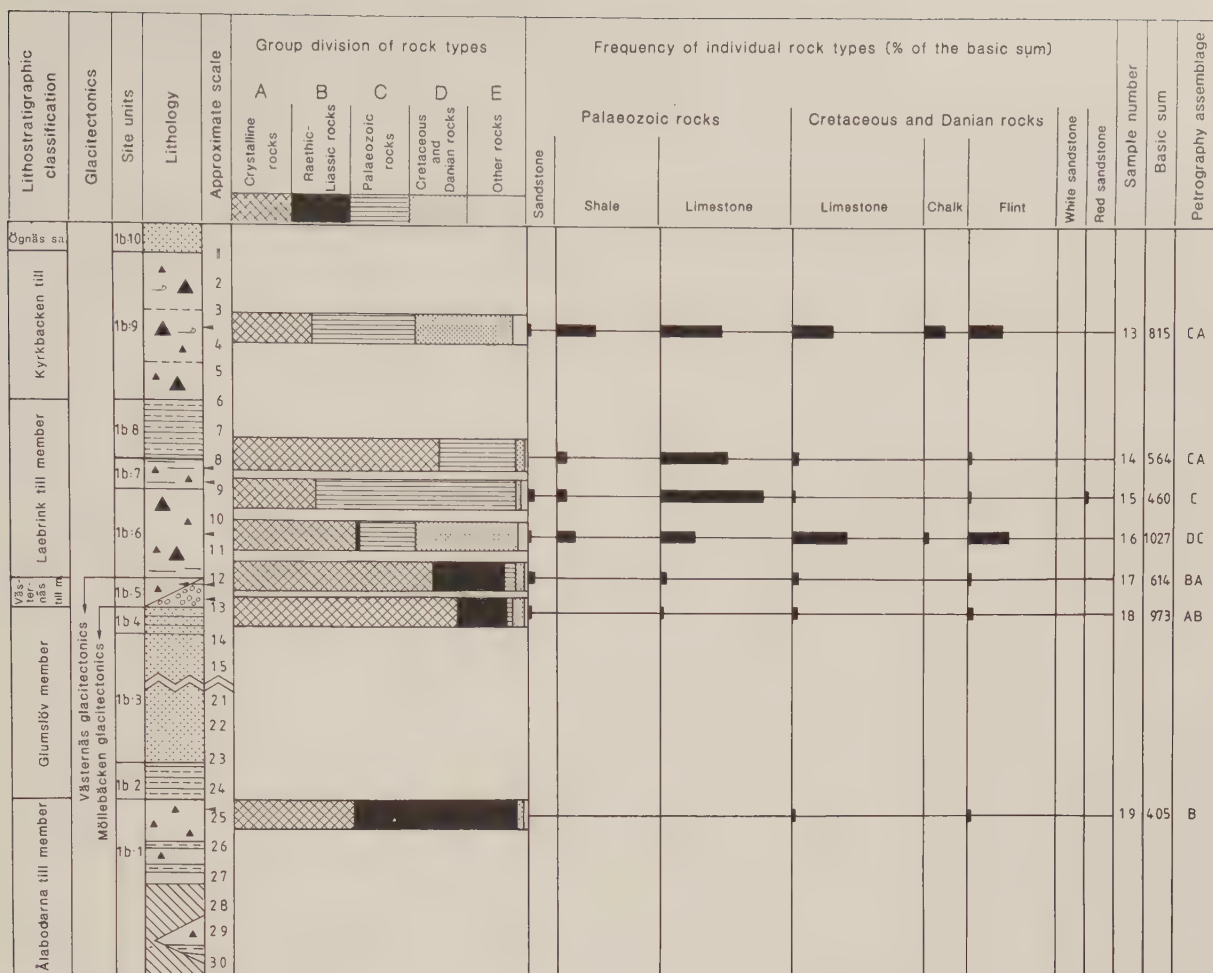


Fig. 7b. Lithostratigraphy and petrography spectra, site 1, part 1:310-620.

stratified sand, at 1:55-65, to gravel and gravelly sand with lenses and thin beds of diamicton at 1:80-90.

Unit 1a:10(Facies E1f). The contact to the underlying unit is sharp. From 1:90 and further to the south-east it rests directly on the dislocated strata. The diamicton has a silt and clay content of approximately 50 %. Some boulders are scattered in the unit. In some places the lower part, 10 to 20 cm thick, contains streaked-out smudges of fine sand and silt. The transition to the homogeneous part is diffuse. One fabric analysis was performed in the upper part of the unit, and the rose diagram shows a marked peak in N80°W/S80°E. A boulder, buried in the substratum at 1:85 and with the upper surface in contact to the fissile diamicton, was striated from S2°W, S20°E and S70°E.

Unit 1a:11(Facies A and B). The unit consists of 10 to 25 cm of stratified sand and gravel at 1:0-90. Further to the south, the unit is traced as a discontinuous gravel bed or a lag of small boulders.

Unit 1a:12(Facies D). The unit consists of crudely laminated silt and clay. It is distinct between 1:55 and 1:90 but difficult to trace further to the south-east.

Unit 1a:13(Facies E2f and E2). The unit consists of a inhomogeneous diamicton with an average silt and clay content of 50 % or

slightly more. The inhomogenities are composed of slabs and lenses of white chalk or red coloured diamicton. Boulders are few and scattered in the unit. A fissility pattern can be seen in some parts of the unit. The fissile parts, facies E2f, and the non-fissile parts, facies E2, are lateral variations within the same unit. One fabric analysis was made in a fissile part of the unit 1 m above the lower contact at 1:75. The rose diagram has a main peak in N30°E/S30°W. The thickness of the unit ranges between 4 and 5 m, and the lower contact is generally distinct.

Unit 1a:14(Facies B). The top unit of the cliff consists of a well sorted fine sand. Dark brown humic layers are frequent in the sand.

Lithostratigraphic units, section 1:310-620

10 units were distinguished, lb:1 to lb:10. lb:1 to lb:5 were affected by deformation, and lb:6 to lb:10 lie as subhorizontal units with an angular unconformity to the lower strata. Unit lb:5 lies as a pocket in the synform of an isoclinal fold in the uppermost part of the deformed sequence and is not involved in the thorough deformation of the lower units. The lower and middle part of the cliff is mainly covered, and the deformation, combined with the scattered outcrops, makes the recording of the middle part of the sequence difficult. A narrow vertical

cut at 1:475 to 1:485 gave a good continuous sequence of the lower strata, and the stratigraphy and the environmental interpretation were based on observations from this outcrop. Scattered observations from other parts of the lower cliff fit well into this stratigraphy.

Unit 1b:1(Facies E3D). The diamicton beds in the stratified facies are texturally heterogeneous and crudely laminated. Some layers contain intraclasts of clay. The intrabeds of silt and clay are laminated. The unit makes up the lower part of the sequence at 1:475 to 1:485 but can also be traced in the lower part of the cliff at 1:420 to 1:450 and at 1:600 to 1:620. The thickness at 1:610 is 6 m, but the unit continues below the present sea level.

Unit 1b:2(Facies D). The transition from unit 1b:1 is gradual. The diamicton beds cease, and the sequence continues as a laminated silt and clay. The alternation between clay laminae and sets of laminated silt is rhythmic.

Unit 1b:3(Facies B). The silt and clay facies grades into sand facies. The mean grain size values lie in very fine sand and coarse silt grades. The primary sedimentary structures are dominated by different types of cross-lamination.

Unit 1b:4(Facies C). The transition from 1b:3 to 1b:4 is transitional. The unit consists of an alternation of sets of cross-laminated sand, interbedded by clay laminae. Two thick beds of convoluted sandy silt occur in the middle of the unit. The unit is intersected by a slickensided shearplane, and the deformation in the middle part of the cliff prevented a further tracing of the unit.

Unit 1b:5(Facies E3B). A pocket of gravel, sand and diamicton is enclosed by sand, probably of unit 1b:4, in an isoclinal fold at 1:315-325. The relative position of the gravel in relation to the diamicton is unclear due to the deformation. Sorted sediments and diamictons were folded together in a pattern that indicates repeated alternations of the sediment types. The boundaries between the layers are distinct, and the strata are interpreted as a folded facies E3B.

Unit 1b:6(Facies E2f-Elf). The diamicton rests discordantly on the deformed units. The lower boundary is sharp as regards structural features, but indistinct as regards the texture. The lower part, 0.1 to 0.5 m thick, consists of strongly sheared sand of the same texture as the substratum. The diamicton content increases gradually upwards, and there is no distinct boundary between facies E2f and facies Elf. The thickness of the unit ranges from 0.5 m to a maximum of 5.5 m in the landslide scar at 1:340. Fabric analyses at 1:340 show a preferred orientation of clast longaxes of N17°W/S17°E at 0.5 m from the base of the unit, N75°W/S75°E at 1.5 m, N78°W/S78°E at 3 m and N70°W/S70°E at 4.75 m. Another fabric analysis at 1:510 gave a preferred orientation of N65°W/S65°E.

Unit 1b:7(Facies E2 and E3B). The upper part of the cliff is also partly covered, and the unit was not traced continuously over the section. Scattered outcrops from 1:450 to 1:510 show facies E3B with diamicton beds of different grain size composition, interlayered and interfingered by sand and gravel, often with a restricted lateral distribution. The petrographic composition is very

heterogeneous with large variations between the individual beds. Exposures from 1:520 to 1:565 show facies E2. The structure in the diamicton is a crude banding, which is mainly seen as distinct colour variations in green, gray and red. The colour banding is accompanied by differences in petrography composition. The individual bands are 2 to 10 cm thick. The two diamicton facies were not seen on top of each other. The thickness of the unit ranges between 0.5 and 1 m.

Unit 1b:8(Facies D). The diamicton is overlaid by a crudely laminated or almost homogeneous clay. The clay contains outsize clasts. The thickness ranges between 0.5 and 2 m.

Unit 1b:9(Facies E2f and E3B+D). A continuous bed of 5 to 7 m of diamicton is found in the upper part of the cliff. The texture as well as the petrography composition is heterogeneous. At 1:470-500 the diamicton is distinctly stratified with silt, clay, fine sand and chalky clay intrabeds. At 1:520-530 the diamicton looks different. It is still very heterogeneous, but there is no distinct stratification. Parts of divergent texture occur as lenses with indistinct boundaries to surrounding diamicton. Smudges, both small and up to 1 m in diameter, occur. One fabric analysis in this part gave a preferred orientation of clast longaxes of N25°E/S25°W.

Unit 1b:10(Facies B). The upper 0.5 m of the cliff consists of a crudely stratified or almost homogeneous medium to fine sand. An indistinct banding of brown humic sand is sometimes present.

Tectonics

The lower 2/3 of the section are affected by tectonics. The deformation is of two types: major faults, as the one at 1:140-170, and recumbent folding, as at 1:320, 1:430, 1:475 and 1:505. The large fault has earlier been described by Johnsson (1962) and Rasmussen (1973a). The vertical displacement of the stratified sediments is 3.5 m, and the thrust plane has a N5°E strike and dips 40° towards the east. The fault continues below the sea level. The original dip of the bedding planes in the laminated silt and clay was probably close to horizontal, and the present increasing inclination towards the west indicates a synclinal bending.

The folding is restricted to the upper part of the deformation and is most intense in the sand at 1:310-320 and 1:420-510. Some fold axes from the destroyed part of the section at 1:310-370 were measured by Rasmussen (1973a, p.27).

My observations of the folding were confined to the south-east part of the section. One overturned fold can be seen at 1:430. The fold axis strikes in N62°W, and the fold is closed in the south-west direction. The core is older than the outer portion. Two recumbent folds have been formed in the sand at 1:475, 13-16 m a.s.l. The lower fold is closed towards the north-east, with the core younger than the outer portion, and the upper fold is closed towards the south-west, with the core older than the outer portion. The strikes of foldaxes are N60°W and N57°W, and the axes lie obliquely to the trend of the cliff. The shear movement during the folding has been along a clay

lamina, and the lower limb of the folds is in a primary position relative to the substratum.

A sand unit with clay laminae in the base has been sheared over the folded sequence. The sand and clay unit passes over into a synclinal bend 20 m further to the south. This bend has an axis in N47°W. The sand inside the bend is intersected by numerous small reverse faults, and the laminae in the front of the bend are wrinkled in small folds.

Some low angle reverse faults were measured in minor cuttings in the lower 5-10m in the south-east part of the section. The measurements are shown in Fig. 8.

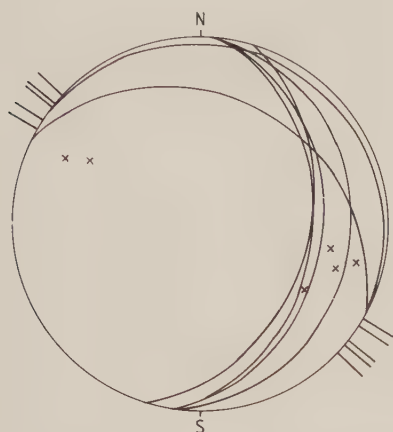


Fig. 8. Tectonics, site 1. Stereographic projection. Thrust faults are plotted as large circles. Bedding is plotted as poles to bedding surfaces on lower hemisphere. The trend of fold axes is shown by marks outside the circle.

5.1.2 Site 2

Site 2 is situated about 300 m to the west of the south cape of Ven. The section (Fig. 9) is 190 m long and trends S80°E. The top of the cliff lies between 20 and 22 m above the present sea level. The lower portion of the cliff in the eastern part is mainly covered by slumps, and only scattered observations were made in small outcrops. Units 2:1 to 2:4 are affected by glaciectonics. The lithostratigraphic log is shown in Fig. 11.

Lithostratigraphic units, section 2:0-190

Unit 2:1(Facies Elf). This unit is exposed in the western part of the section. The diamicton is homogeneous with a high silt and clay content and a low ratio of clasts of gravel size and larger. The diamicton has a widely spaced fissility pattern. One fabric analysis was made 1 m above the sea level at 2:20-25. The longaxes of the clasts have a preferred orientation of N10°W/S10°E.

Unit 2:2(Facies D). The diamicton at 2:95-100 is overlaid by laminated silt and clay.

Unit 2:3(Facies B). The silt and clay facies grades into sand facies. The thickness of the unit was not possible to estimate, as the strata are dislocated and mainly covered.

A small exposure 1.5 m further up in the cliff shows cross-laminated and planar laminated sand.

Unit 2:4(Facies D). Minor outcrops with sand, silt and clay occur in the lower part of the cliff along the section between 2:70 and 2:190. The silt and clay at 2:105-120 has a substratum of sand. This sand was correlated to the unit 2:3 at 2:95-100.

Unit 2:5(Facies Elf). This unit lies discordantly on the lower dislocated strata. The unit increases in thickness towards the west and has a maximum thickness of 4.5 m. The diamicton consists of approximately up to 50% of silt and clay. The content of gravels, cobbles and boulders is moderate, but compared to other diamictons in the area it is high. Two fabric analyses were performed in the diamicton, one in the lower part and one in the upper part of the unit. The former had a preferred orientation of clast longaxes at N30°E/S30°W and the latter at N45°E/S45°W. A buried boulder in the boundary to the 2:6 unit had two systems of striae. The part of the boulder that was buried in the lower unit was striated from N50°E, while the surface in contact to the upper diamicton had a system of very fine striae with a S10°W to S8°E direction.

Unit 2:6(Facies Elf). The contact between the two diamictons is very sharp. Unit 2:6 has a slightly higher clay content than the lower diamicton. The petrography content differs significantly, and the sharp contact is best seen as a colour difference and a difference in the petrography of the clasts. Scattered boulders occur in the diamicton. Striated and flat-iron-shaped clasts are frequent. The clast longaxes, 0.5 m above the base, have a preferred orientation at N10°W/S10°E. In the middle of the 7 m thick unit the preferred orientation is N35°W/S35°E.

Unit 2:7(Facies A and E3B). The diamicton is covered by a subhorizontal unit of pebbly gravel, which gradually changes into stratified diamicton with sand and gravel towards the east. The unit is 0.6 to 1.1 m thick.

Unit 2:8(Facies D). The contact to the lower gravel and stratified diamicton is fairly sharp. The unit has a thickness of 4.5 m, and the lower part is dominated by silt with convolute lamination. The clay content increases upwards.

Unit 2:9(Facies Elf). The contact to the lower unit is sharp. Small low-angle edges of clay were sheared into the base of the diamicton. The lowermost part of the diamicton has a very high clay content. The main part of the unit is quite homogeneous with a closely spaced fissility pattern. A fabric analysis, 1 m from the base of the unit, gave a preferred orientation of clast longaxes at N10°E/S10°W.

Unit 2:10(Facies B). The upper 0.5 m consists of structureless fine to medium sand.

Tectonics

The lower units 2:1 to 2:4, are affected by tectonism. A major shear-plane was developed in the diamicton at 2:90, and some minor overthrusts and angular folds were observed in the sand at 2:95. The measurements are plotted in the stereographic projection (Fig. 12).

Site 2, section 2.0-190

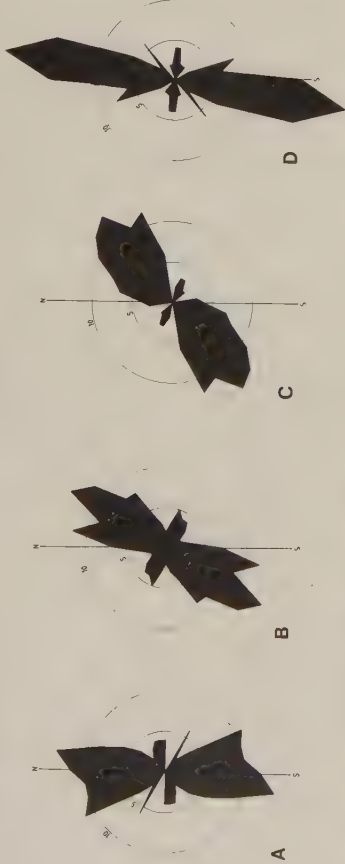
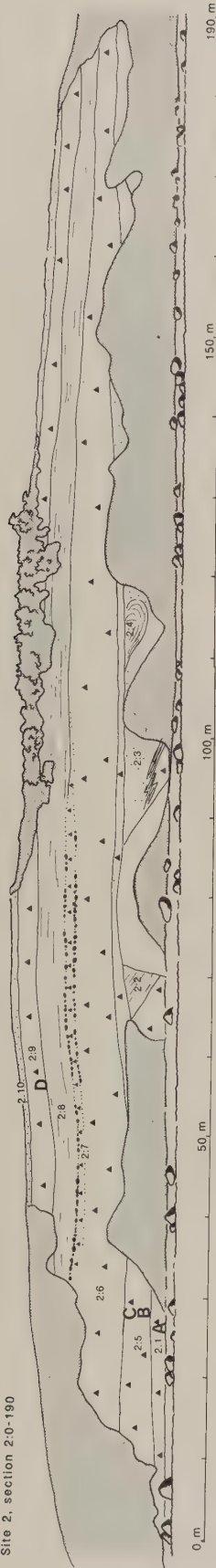


Fig. 9. Section chart, site 2. The rose diagram show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

Site 3, section 3.0-45

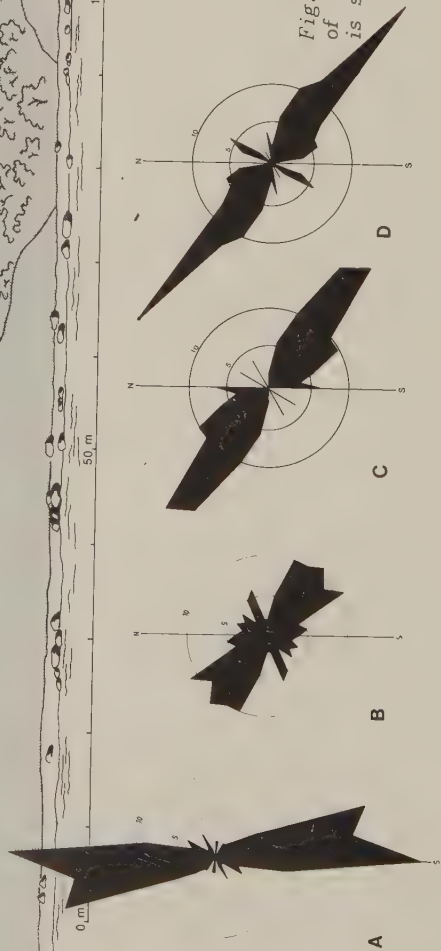
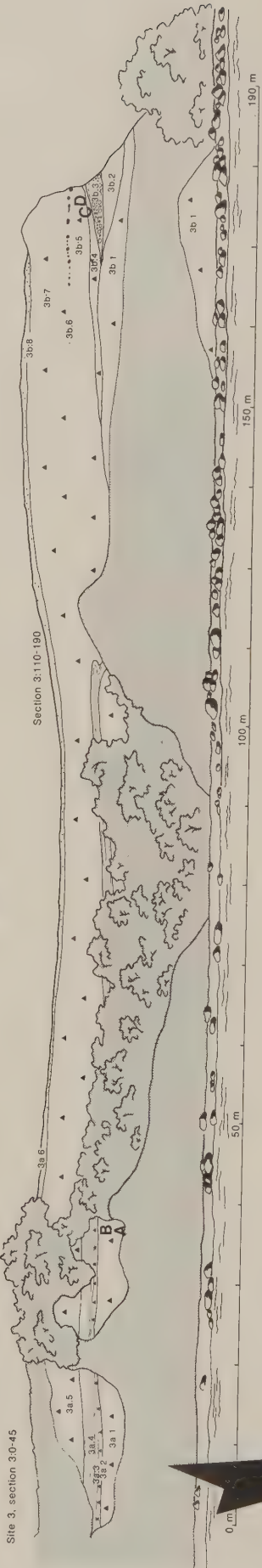


Fig. 10. Section chart, site 3. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

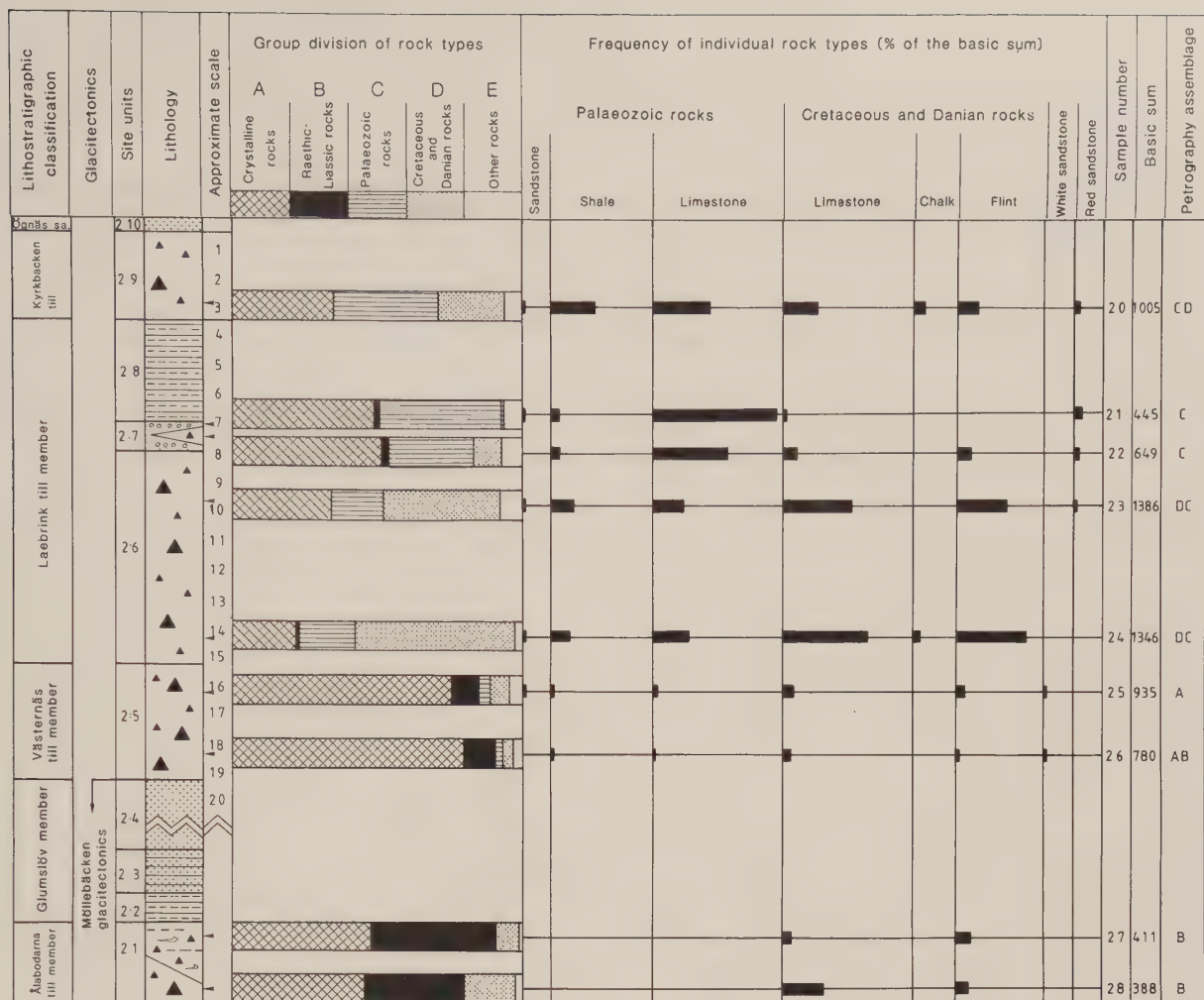


Fig. 11. Lithostratigraphy and petrography spectra, site 2.

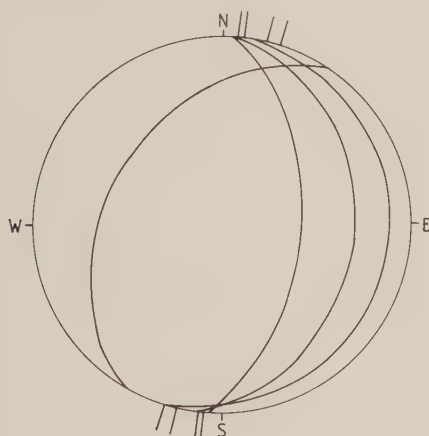


Fig. 12. Tectonics, site 2. Stereographic projection. Thrust faults are plotted as large circles. The trend of fold axes is shown by marks outside the circle.

5.1.3 Site 3

Site 3 (Fig.10) extends from the south cape of Ven and 200 m westwards. The cliff has a general trend of S74°E. Most of the strata in the cliff are covered, and the main information was collected from the upper, not

dislocated strata. Two stratigraphic records were made, one for the 3:0-45 portion (Fig.13) and one for the 3:110-190 portion (Fig.14). The upper two units in each record link the strata in the western part to the strata in the eastern part.

Lithostratigraphic units, section 3:0-45

Unit 3a:1 (Facies E1f). The unit is exposed in the middle part of the cliff between 3:0 and 3:45. The substratum was not exposed, but post auger drilling indicates that silt and clay are hidden behind the slid masses. The thickness of the diamicton is approximately 4 m. The silt and clay content is high in the homogeneous diamicton and ranges between 50 and 60%. Boulders are scattered in the outcrop. Fabric analyses were performed in the lower and upper parts of the unit at 3:35. The preferred orientation of clast longaxes in the lower part is N10°W/S10°E, and close to the top N50°W/S50°E.

Unit 3a:2 (Facies E3B). Thin, lens- and wedge-formed beds, 1 to 10 cm thick, interlayered by sand and gravel, lie on the fissile diamicton at 3:15-20. The petrography composition is similar to the lower unit. The thickness of the unit ranges from a few centimeters to 0.6 m.

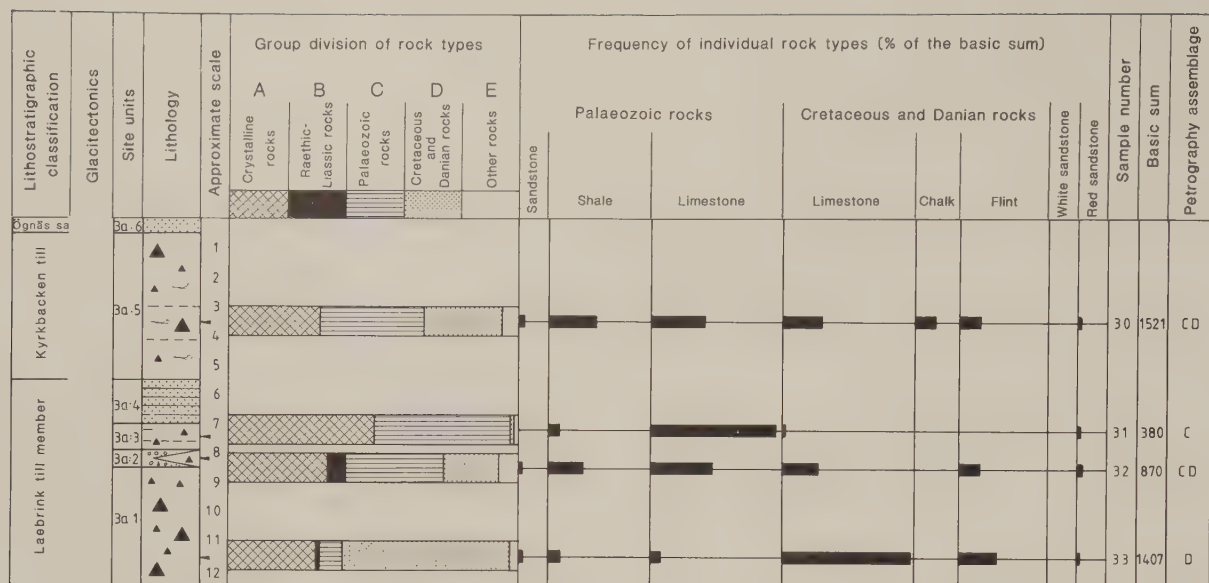


Fig. 13. Lithostratigraphy and petrography spectra, site 3, part 3:0-45.

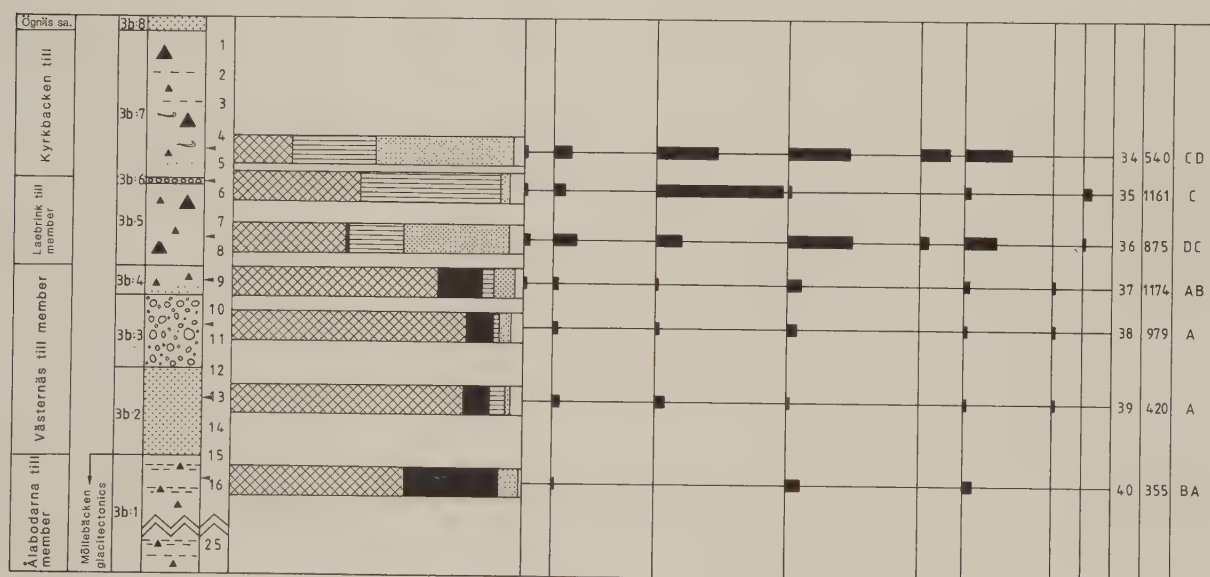


Fig. 14. Lithostratigraphy and petrography spectra, site 3, part 3:110-190.

Unit 3a:3(Facies E3D). The character of the diamicton changes upwards. At 3:15-20 the texture of the intrabeds changes into silt and clay. This change can be traced along the entire section. At the same time the petrographic composition changes with a remarkable increase in the ratio of paleozoic limestone. Slump structures, folding and shearing are present in the upper part of the unit, which is approximately 0.9 to 1 m thick.

Unit 3a:4(Facies C). Undisturbed strata of cross-laminated and parallel-laminated very fine sand and silt lie on top of the disturbed lower sediments. The unit is 1.5 m thick at 3:15 - 20.

Unit 3a:5(Facies E2). The unit contains extensive, thin bands or layers of white chalk. Smudges of red-coloured diamicton with a high content of reddish sandstone are also

present. Boulders are rare compared to unit 3a:1. The unit has an average thickness of 5-6 m.

Unit 3a:6(Facies B). A thin layer of structureless humic sand covers the sequence.

Lithostratigraphic units, section 3:110-190

Unit 3b:1(Facies E3D). Several small exposures were made by digging in the lower part of the cliff. All of the exposures contain diamicton, silt and clay. The diamicton and silt layers often have intraclasts of clay. The diamictons are faintly stratified, cobbles and boulders are very rare, and the silt and clay content is high. Bedding planes in the stratified sequence have a general dip towards the east.

Unit 3b:2(Facies B). The unit lies with an angular unconformity on the stratified diamicton in a broad channel or trough just at the south cape. Primary sedimentary structures are dominated by planar lamination with subhorizontal bedding planes. Trough cross-bedding with gravel lags and poorly sorted gravel residual also occur in the sand.

Unit 3b:3(Facies A). The gravel lies with a sharp erosive contact on top of the sand. The unit is also restricted to the channel at the south cape. The gravel is poorly sorted with a high content of cobbles and boulders. There is no obvious bedding in the gravel, but a fabric analysis gave a well preferred orientation of clast longaxes at N50°W/S50°E.

Unit 3b:4(Facies E3B). The diamicton on top of the gravel is distinctly stratified with layers of sand and gravel between thin layers of sandy diamicton. The unit is 0.7 m thick and seems to have the same lateral distribution as the sand and the gravel, i.e. it is restricted to the trough.

Unit 3b:5(Facies E1f). A distinct lower boundary of the diamicton unit was traced from 3:145 to 3:180. A sharp upper boundary was also traced along the section. The thickness of the diamicton bed between the distinct boundaries is approximately 3 m. A close examination of the diamicton bed shows, however, that the unit can be divided into two parts. The lower part, 0.2 to 0.7 m. thick, differs from the upper part in petrographic composition. A comparison to lower units, 3b:3, 3b:4 and 4a:6, shows that the petrographic composition in the lower part of unit 3b:4 is equal to these units. The transition from one petrographic type to another is very indistinct, and it was not possible to distinguish the unit on the basis of structural evidences. Thus, the diamicton is considered one single unit with a gradual vertical change in petrographic composition. Fabric analyses in the two parts confirm this interpretation. The rose diagram from the lower part has a main peak in N55°W/S55°E and from the upper part in N50°W/S50°E.

Unit 3b:6(Facies A). The upper surface of the diamicton is partly covered by a discontinuous bed of gravel, generally only 0.1 to 0.2 m thick. The unit is more continuous on the eastern side of the cape (unit 4a:10).

Unit 3b:7(Facies E2). Equal to unit 3a:5.

Unit 3b:8(Facies B). Equal to unit 3a:6.

5.1.4 Site 4

Site 4 (Fig.15) extends from the south cape and 400 m eastwards. The orientation of the cliff varies considerably due to two semi-circular landslide scars at 4:90-150 and 4:220-300. The general trend of the coastline is N70°E.

The lithostratigraphy is shown in two sequences (Figs 16 and 17). The western part of the section, 4:0-140, is similar to the eastern part of section 3, and most of the units can be physically traced from one site to the other. The eastern part of the section, 4:140-400, is dominated by the dislocated strata in the lower part of the sequence.

Lithostratigraphic units, section 4:0-140

Unit 4a:1(Facies E3D). Observations were made in minor outcrops and excavations in the lower and middle part of the section. The lower boundary was not visible in the cliff, and the unit extends below the present sea level. The upper boundary is sharp and can be studied in the thrust part at 4:25 and in the basal part of the cliff at 4:90. The diamicton layers dominate in the exposures. The layers have high silt and clay content. Pebbles and boulders are extremely rare. The diamicton beds often show an internal structure of crude lamination. Intraclasts of clay are present in some of the layers. The diamicton beds are interlayered by clay laminae or more complex sets of cross-laminated and parallel-laminated silt. The individual intrabeds seldom exceed 20 cm in thickness.

Unit 4a:2(Facies B). The diamicton is overlaid by sets of cross-laminated fine to very fine sand. The sand has a thickness of 0.6 m at 4:25 and 2 m at 4:90.

Unit 4a:3(Facies C). At 4:90 the sand grades into 0.7 m of alternating bed facies with cross-laminated sand sets and clay laminae.

Unit 4a:4(Facies D). 1.5 m of laminated silt and clay was exposed on top of the 4a:3 unit at 4:90. Higher up the sediments were covered by slumps. Post auger drilling was made through the slump masses, and the silt and clay facies was further traced for 1.5 m.

Unit 4a:5(Facies B). The drillings indicate another sand bed on top of unit 4a:4. The same stratigraphy, 4a:2 to 4a:5, was interpreted from outcrops in the middle of the cliff between 4:60 and 4:70. The outcrops are, however, small, and the strata were not continuously traced bed by bed.

Unit 4a:6(Facies B). Stratified sand lies with an angular unconformity on the lower deformed strata at 4:0-50. The unit has a maximum thickness of 6 m. The sand is predominantly a planar-laminated medium to coarse sand, but it also contains thin subhorizontal layers of gravel and sandy diamicton.

Unit 4a:7(Facies A). The coarse gravel is restricted to the south cape and has been described as unit 3b:3. The three-dimensional extension of the sand and gravel body can be indirectly interpreted from the continuous leakage of ground water from the sand in the cliff. It is likely that the body extends towards the interior of the island and drains a minor area.

Unit 4a:8(Facies E1f-E2f). The sand and the gravel at the south cape are covered by a thin unit of diamicton, which increases in thickness to 5-6 m at 4:60. The thickness gradually decreases towards the east, and the unit ends at 4:140. The contact to the stratified sediments of the substratum is mainly a sheared zone of mixing (facies E2f). Primary sedimentary structures in the lower sediments are completely destroyed, and an 0.1 to 0.3 m thick zone of crude stratification of diamicton and sorted sediments, often with hook- or z-formed folds, constitutes the lower part of the diamicton unit. The main part of the diamicton is homo-



section 4:140-400

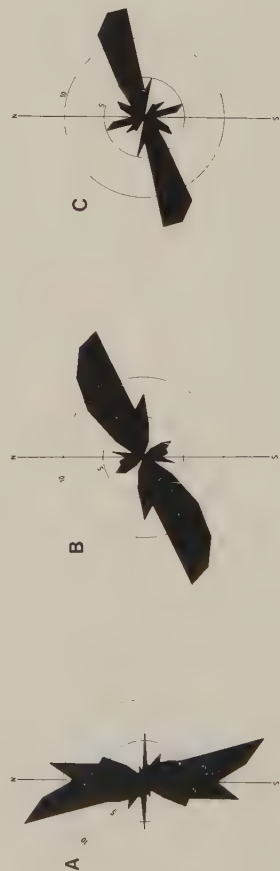
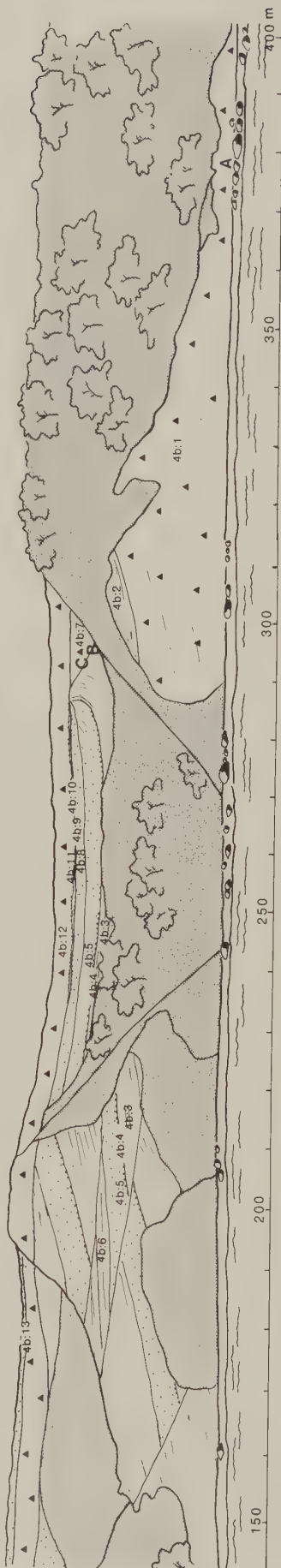


Fig. 15. Section chart, site 4. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

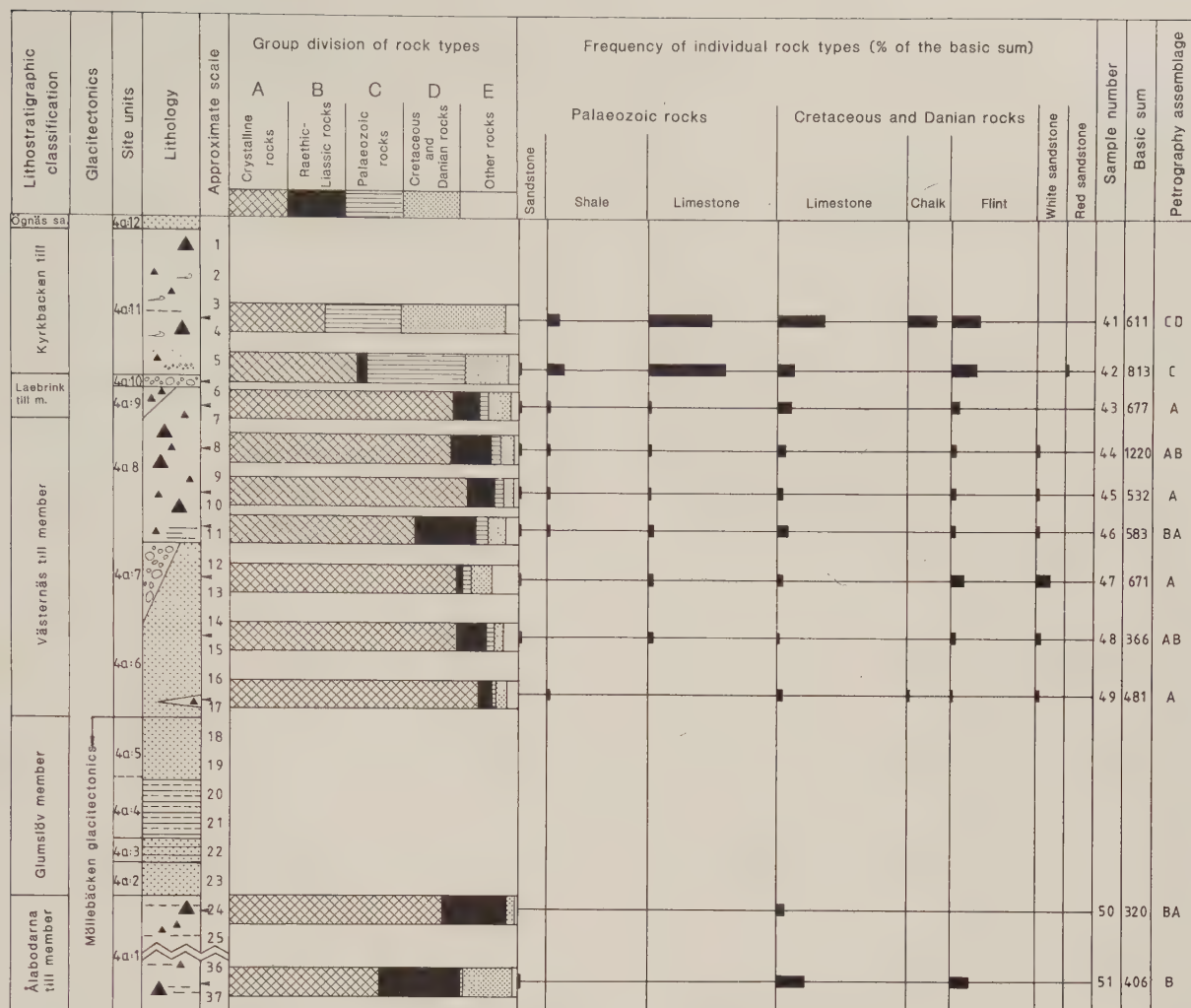


Fig. 16. Lithostratigraphy and petrography, site 4, part 4:0-140.

geneous, the silt and clay content is moderate, and boulders are scattered in the unit. Striae were measured on some of the boulders and showed directions of N25°E, N40°E, N60°E and N40°E, and N35°E.

Unit 4a:9(Facies E1f). Unit 4a:8 is overlaid by another fissile diamict. This diamict is the 3b:5 unit of section 3. In section 4 it extends from the south cape to 4:30 as a wedge-formed bed. The upper part of the cliff is an almost vertical wall, and the contact between unit 4a:8 and 4a:9 was only studied at a distance. For further description see unit 3b:5.

Unit 4a:10(Facies A). The unit was studied at a distance and was recognized as layers of gravel and sand, which in some parts are replaced by a concentration of pebbles and boulders.

Unit 4a:11(Facies E3B-D). The unit has a uniform thickness of approximately 5 m and covers the whole section. Fresh cuttings after a landslide at 4:110-140 showed a distinct stratification with diamict beds of various texture. The thickness of the beds ranges from some centimeters to approximately 1 m. The geometry of the beds varies from restricted lenses to wedge-formed beds and more extensive tabular beds. Trough-formed

erosive contacts and deformation structures are frequent.

Unit 4a:12(Facies B). A thin discontinuous unit of structureless sand with humus content lies on the top.

Lithostratigraphic units, section 4:140-400

The main part of the sediments is dislocated and intersected by low-angle shear planes. The general dip of the bedding-planes is towards the west, and the lowest unit was found in the eastern part of the section between 4:285 and 4:400. The sequence is divided into 13 lithostratigraphic units, 4b:1 to 4b:13, of which the lower 6 units have been dislocated.

Unit 4b:1(Facies E2f and E3D). The unit is exposed in a high vertical wall at 4:285-400. The wall has a maximum height of 17 m at 4:300 and declines towards the east. The diamict in the wall is faintly laminated with a light coloured banding of coarse silt. The lateral extent of those bands is generally less than 2 m. The lower part of the unit at 4:375 to 4:400 has a widely spaced sub-horizontal fissility pattern. Columnar jointings also occur in the lower part. The silty diamict has an extremely low content of cobbles

Unit 4b:7(Facies Elf). This unit lies with an angular unconformity to the lower deformed units in the eastern part of the landslide scar. The exposure is almost perpendicular to the coast line, and the section chart (Fig. 15) shows only a very small part of the cutting. The thickness of the unit is 5.5 m in the cutting but decreases to zero in the inner part of the semicircular scar, and the unit is not to be found in the western part. The diamicton is very homogeneous, and the silt and clay content is around 50%. Boulders are scattered in the wall, and some of them are striated on bevelled surfaces. One boulder, situated 1 m above the base of the unit, was striated from N12°E. Another boulder at the same level had two striae systems, one from N20°E and one from N58°E. Two boulders in the middle of the unit were striated from N60°E and N64°E. One boulder in the upper part was striated from N70-75°E. Two fabric analyses from the middle and upper part of the unit showed preferred orientation of clast longaxes at N60°E/S60°W and N70°E/S70°W.

Unit 4b:8(Facies B). This unit fills up a shallow depression between 4:215 and 4:280. The sand contains thin diamicton layers of restricted lateral extent. It also contains lenses of gravel and silt. The heterogeneous unit has a thickness of approximately 2 m. The sand and the next three units were only traced in the inner wall of the landslide scar.

Unit 4b:9(Facies A). The sand is covered and partly truncated by a horizontal bed of gravel. The gravel is 0.2 m thick.

Unit 4b:10(Facies D). The unit is 0.1 m thick

and laterally extensive and consists of laminated silty clay.

Unit 4b:11(Facies A). The upper surface of the silty clay is scoured with small channels or troughs. The troughs are infilled with gravel. The unit is discontinuous.

Unit 4b:12(Facies E2 and E3). Unit 4a:11 was directly traced into unit 4b:12, which continues with equal thickness of approximately 5 m in the eastern part of section 4. The stratified facies continues at least to 4:215, but the two facies, distinctly stratified and crudely stratified or smudged facies, are difficult to distinguish in the vertical inaccessible cliff-walls. At 4:260 to 300 there is no distinct stratification. The diamicton is, however, very inhomogeneous with white smudges of chalk, some of them up to 0.5 - 1 m² large, and lenses or smudges of red-coloured diamicton.

Unit 4b:13(Facies B). The humic sand on the top is discontinuous and has a maximum thickness of 0.5 m.

Tectonics

The tectonics is best seen at 4:150-250, where the sediments are intersected by a series of low-angle shear-planes (Figs 18 and 19). Asymmetrical drag folds have been formed on the flanks of the shear-planes. Ripple heights in the cross-laminated silt and sand have been exaggerated by compression.

The total section gives the impression of being a large basin-like flexure. The general dip of bedding-planes is towards the east in the western part and towards the west



Fig. 18. The dislocated strata, units 4b:3 to 4b:6, at 4:150-250. The cliff trends approximately west and the blocks are thrust from the east.

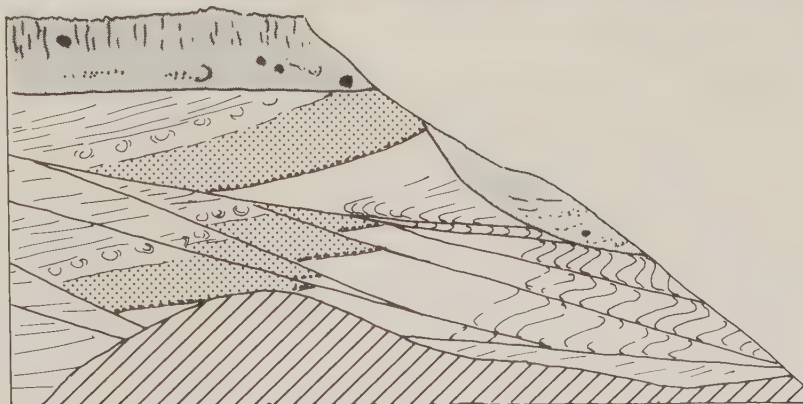


Fig. 19. A sketch, showing the position of shear planes at 4:150-250 (compare with Fig. 18).

in the eastern part. The upper boundary of the lower diamicton (4a:1, 4b:1) lies at 20 m a.s.l. in the west, declines and disappears below the sea level in the middle part, and can be found again at 17 m a.s.l. in the eastern part. The flexure is transversed by shear faults with a general dip towards the east. At 4:20-25 there are two high-angle reverse faults with a N 2-5°E strike. The faults converge upwards, and the two outward blocks have moved upwards in relation to the middle block. Cataclasis has resulted in a 4-5 cm wide zone of crude foliation along the faults.

Dips of bedding-planes, shear-planes and axes of drag flexures from site 4 are shown in Fig. 20.

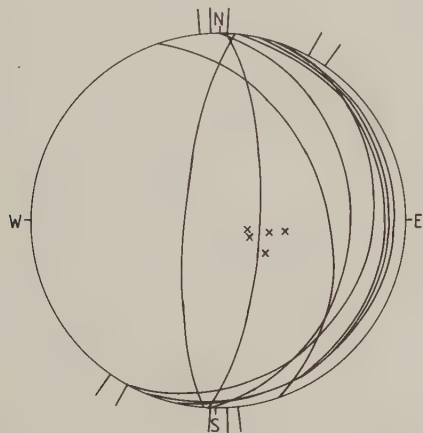


Fig. 20. Tectonics, site 4. Stereographic projection. Thrust faults are plotted as large circles. Bedding is plotted as poles to bedding surfaces on lower hemisphere. The trend of fold axes is shown by marks outside the circle.

5.1.5 Site 5

Site 5 (Fig. 21) is situated at the south-east cape of the island. The cliff trends N85°E and the measured section is 110 m long. There are two main outcrops, one in the lower western part of the cliff and one in the lower eastern part. The upper portion of the cliff and the space between the two outcrops are mainly covered. An almost sub-horizontal shearplane intersects the section, and the lower units of the recorded stratigraphy are found in the hanging wall above the stratigraphic units in the footwall. The compiled lithostratigraphy is displayed in Fig. 22.

Site 5, section 5:0-110

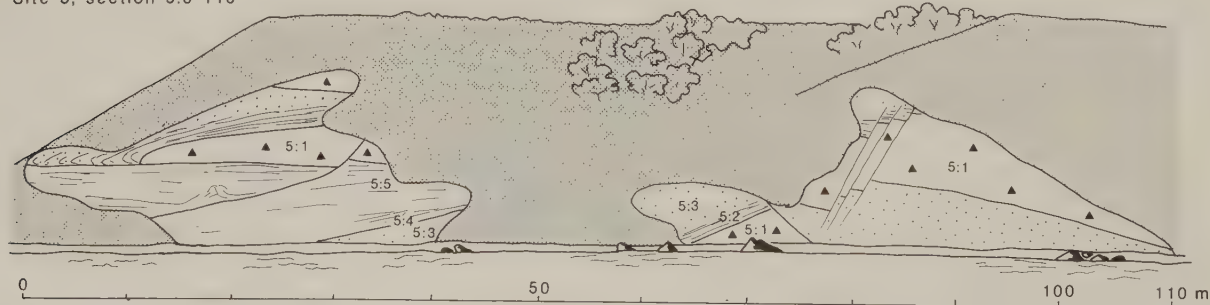


Fig. 21. Section chart, site 5.

Lithostratigraphic units, section 5:0-110

Unit 5:1 (Facies E2 and E3D). The lowest unit in the hanging wall consists of diamicton. A shear plane intersects the unit obliquely and the thickness of the unit increases from 0.5 m at 5:10 to 7 m at 5:80-90. The diamicton has a very high silt and clay content. Pebbles and boulders are rare. The diamicton is crudely to distinctly stratified. Some parts of the 7 m high exposure at 5:85-105 can be divided into 0.2 to 1 m thick beds of homogeneous diamicton interbedded by crudely laminated diamicton beds and beds with intraclasts of clay. Silt and clay facies is subordinate, but thin composite sets of laminated silt and clay are dispersed in the whole unit.

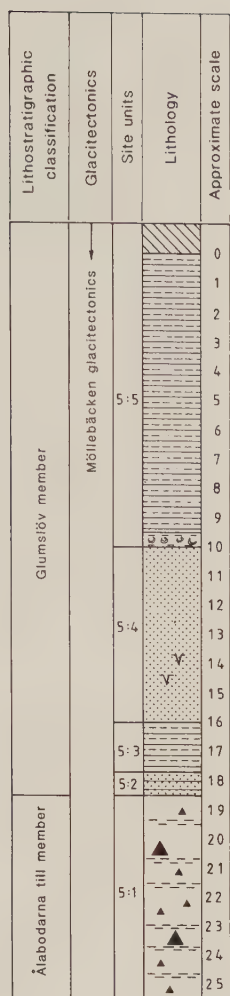
Unit 5:2 (Facies C). The diamicton at 5:10-30 is overlaid by 0.8 m of alternating bed facies with distinct clay laminae, separated by single or multiple sets of cross-laminated very fine sand.

Unit 5:3 (Facies D). The alternating bed facies grades into 1.7 m of laminated silt and clay at 5:10-30. At 5:85-105 there is a direct transition from stratified diamicton to laminated silt and clay.

Unit 5:4 (Facies B). The strata at 5:20-35 were traced into cross-laminated and planar-laminated sand. Approximately 3.5 m of sand was exposed, but the unit was then covered. The same transition from facies D to facies B was also found at 5:80-85. The sand at 5:70-85 is also dominated by cross-lamination and planar-lamination. Two ice-wedge casts were found in the sand. The bedding-planes in the sand dip slightly towards the west. The unit could not be continuously traced between 5:50-70, but post auger drilling in the covered portion indicated a continuous sand unit to 5:45-50 where the upper boundary of a sand unit was exposed. The total thickness of the sand unit is estimated to approximately 5-6 m.

Unit 5:5 (Facies D). The transition from the sand to the silt and clay facies is through a bed of deformed and partly dissolved sandy silt. The bed is 0.3 to 1.0 m thick. The silt and clay facies has a thickness of 10 m, but the upper part is bounded by the subhorizontal shear plane. The sediments show a wide range of structures with parallel-lamination, different types of cross-lamination, and thick units of deformed, dissolved and convoluted bedding. The stratigraphy of the upper part of the slope has not been investigated. The outcrops are too small

Fig. 22. Lithostratigraphy, site 5.



and widely spaced, and it was difficult to trace the lateral and vertical extent of individual units.

Tectonics

Two types of tectonics are present. The sub-horizontal shear-plane, which already has been mentioned, has a slight dip of 0-10° towards the east. Linear slickenside structures were measured on the shear planes. The trend ranges between N 88°E and S 80°E. Drag folds have been developed on the upper flank of the subhorizontal shear plane. The fold axes are in N 2-10°E. The thrust units have later been intersected by some normal faults. The fault at 5:20-30 has a curved slip surface. A series of high-angle normal faults at 5:85-90 has a repeated down-throw in a S50°W direction. The tectonics is shown in Fig. 23.

5.1.6 Site 6

The site is situated in the northern part of the island, about 1 km to the east of the north-west cape. The section (Fig. 24) is 200 m long and trends S75°W. The cliff is 15 m high in the eastern part and declines to 7 m in the western part. The upper portion of the cliff is almost vertical, and the diamictons, on top of the lower deformed strata, were only studied in the lower accessible part. Lithostratigraphy and gravel analysis results are shown in Fig. 25.

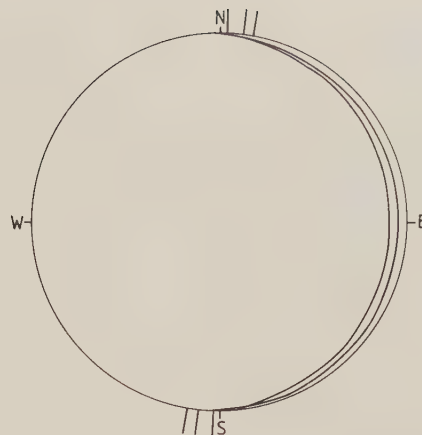


Fig. 23. Tectonics, site 5. Stereographic projection. Thrust faults are plotted as large circles. The trend of fold axes is shown by marks outside the circle.

Lithostratigraphic units, section 6:0-200

Unit 6:1(Facies B and D). Outcrops in the mainly covered lower part of the cliff between 6:85 and 6:200 show thrust strata of fine sand, silt and clay. The sediments were not closely investigated, and the restricted outcrops did not permit any stratigraphic subdivision of the sediments.

Unit 6:2(Facies B). The contact between unit 6:1 and 6:2 was studied at 6:85-100. The boundary is very sharp, and the upper unit truncates the bedding of the lower unit. The bedding planes of the upper unit are parallel to the contact surface, which dips towards the east. The lower part of unit 6:2 consists mainly of planar- and cross-bedded sand with a minor content of granules and pebbles. The sand becomes finer upwards, and cross-lamination becomes the most abundant sedimentary structure. Laminated silt also appears in the upper part. The thickness of the unit is 4.3 m.

Unit 6:3(Facies Elp). This unit is the diamicton in the lower part of the cliff at 6:0-65. The upper boundary to the 6:5 unit is knife sharp. The lower boundary lies below the sea level. The diamicton is homogeneous, but subhorizontal sand partings, 0.1 to 0.3 m apart, were continuously traced over the whole cutting. A fabric analysis gave a preferred orientation of clast longaxes at N25°W/S25°E.

Unit 6:4(Facies A). A channel infilling of gravel can be seen at 6:65-70. The eastern side of the channel consists of diamicton (unit 6:3), and the western side of the channel consists of sand (unit 6:2). The gravel is poorly sorted. Pebbles and boulders are frequent. Intraclasts of silt, clay and also sand occur in the gravel.

Unit 6:5(Facies E2f-E1f). The diamicton lies with an angular unconformity to the dislocated strata in the western part of the section, but the boundary is parallel to the partings in the lower till in the eastern part. The lower part of the diamicton is crudely laminated with thin structureless bands or smudges of silt and sand. The diamicton gradually turns into homogeneous diamicton. The banded zone is sheared, and

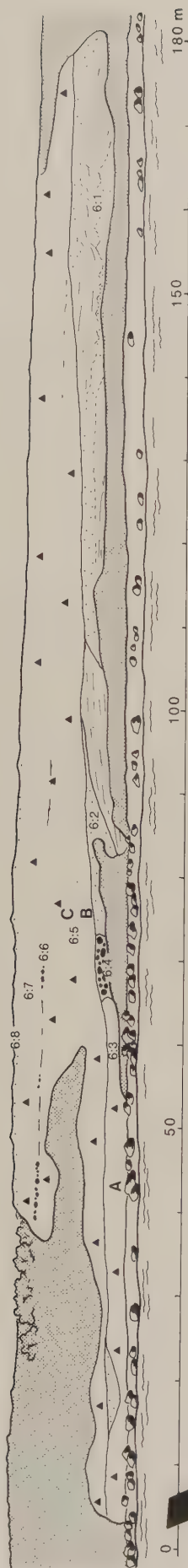


Fig. 24. Section chart, site 6. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

hook-formed folds are present. The homogeneous diamicton has a silt and clay content of around 50%. Cobbles and boulders are scattered in the unit. Two fabric analyses were made at 6:75. The preferred orientation of clast longaxes at 0.5 m from the base of the unit was N5°W/S5°E, and at 2 m from the base N40°W/S40°E.

Unit 6:6(Facies A). The strata in the upper inaccessible part were only studied from a distance. In the eastern part of the section it was possible to distinguish a sub-horizontal layer of gravel, which divided the diamicton into two beds. The gravel disappears westwards, and the boundary between the two diamictons becomes more diffuse.

Unit 6:7(Facies E2). The upper diamicton in the vertical wall is more inhomogeneous than the lower diamicton (unit 6:5). Smudges of different colour and texture are present.

Unit 6:8(Facies B). The top soil consists of 0.5 m of sand mixed with humus.

Tectonics

The tectonic structures in the section were poorly investigated and mainly with regard to the stratigraphic interpretation. The contact between the 6:1 and 6:2 units can be studied at 6:85-100. The lower unit is affected by tectonism and intersected by numerous minor faults. The tectonic features are cut by the erosive surface between the units, and thus the lower sediments were dislocated before the deposition of the second unit. The bedding in the 6:2 unit shows that the sand is not in a primary position. The contact surface and the bedding dip 15-25° towards N70°E. The sand partings of the diamicton unit 6:3 are, however, sub-horizontal.

5.1.7 Site 7

The site is the low cliff, Læbrink, in the north-west part of the north coast. The investigated section (Fig. 29) is 240 m long and strikes in S85°W. The top of the cliff lies 7 m a.s.l. Only the upper 2-3 m are uncovered. Temporary outcrops are exposed in the lower part by wave action during winter storms and show intense deformation of the lower strata in the middle and eastern part of the section. The reconstruction of tectonic features and the recording of stratigraphic units in this lower part are difficult due to the limited outcrops. The stratigraphy is shown in Fig. 26.

Lithostratigraphic units, section 7:0-240

Unit 7:1(Facies Elp). Small outcrops of diamicton are present in the basal part of the cliff at 7:80-90, 7:105-120 and 7:145-170. The diamicton cannot be physically traced from one outcrop to another, but all outcrops show the same type of homogeneous diamicton with sand partings.

Unit 7:2(Facies C). The diamicton at 7:105-120 is overlaid by 0.3 m of parallel-laminated and cross-laminated silt and fine sand, which grades into the next unit.

Unit 7:3(Facies B). The sand sequence is 2 m thick and coarsens upwards. Thin layers

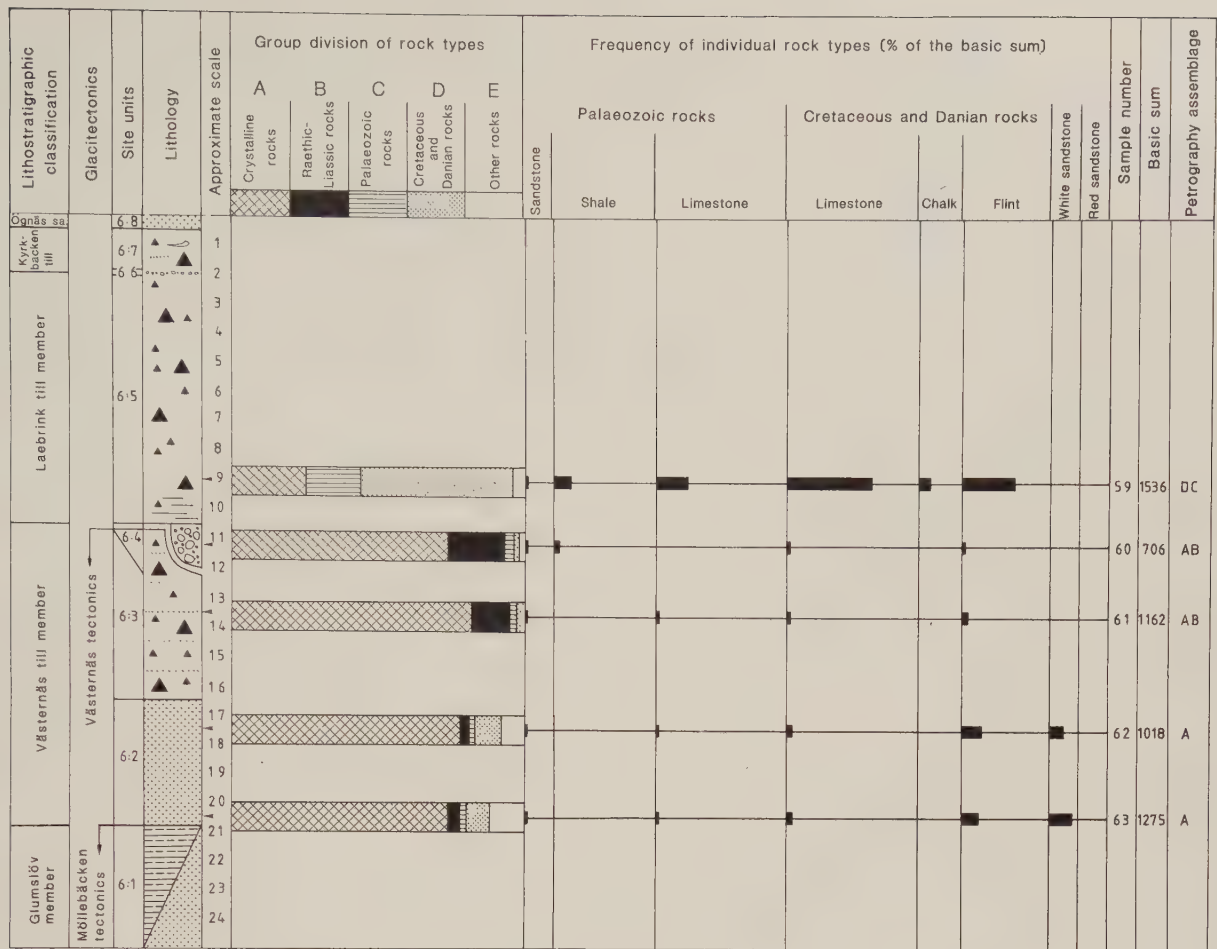


Fig. 25. Lithostratigraphy and petrography spectra, site 6.

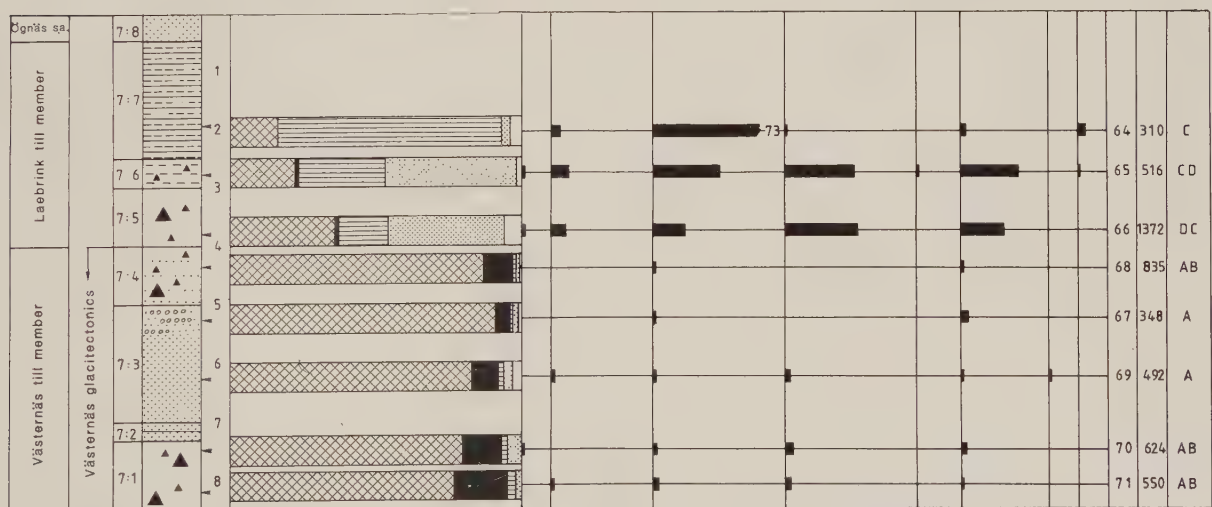


Fig. 26. Lithostratigraphy and petrography spectra, site 7.

of gravel, mainly of granule size, are frequent in the upper part. The silt and sand facies at 7:105-120 were correlated to a similar sequence at 7:0-25, where 0.4 m of laminated silt and clay is exposed at the base of the cliff. The upper surface of the unit is slightly scoured and overlaid by sand, containing thin gravel beds.

Unit 7:4(Facies E3B). The upper part of the 2 m thick sand sequence at 7:15-25 contains

thin diamicton beds. The ratio of diamicton beds increases, and the facies changes to stratified diamicton with sand and gravel intrabeds. The stratified diamicton is at least 1 m thick and is truncated by the erosive surface that separates the lower deformed units from the upper subhorizontal units.

Unit 7:5(Facies Elf). The lower unit of the subhorizontal strata consists of a thin bed,

0.2 to 1.0 m thick, of fissile diamicton. The lower contact is sharp with an undulating course. The trough-shaped downbendings in the subsurface continue as grooves which trend S-S10°W. Fabric analyses in the fissile diamicton at 7:0-5 and 7:190 give a preferred orientation of clast longaxes at N/S.

Unit 7:6(Facies E3D). A stratified diamicton unit can be traced on top of the fissile facies along the entire section. The individual diamicton layers are generally thin, 1 to 15 cm, and are interlayered by silt and clay laminae and cross-laminated very fine sand. A few lenses of gravel were also present in the unit. The thickness of the stratified diamicton is generally less than 0.5 m.

Unit 7:7(Facies D). The top of the sequence is a homogeneous or faintly stratified clay with dispersed outside clasts. The clay content is high (> 50%). The thickness of the stratified diamicton is generally less than 0.5 m.

Unit 7:8(Facies B). An 0.2-0.5 m thick unit of sand covers the strata in the cliff. The unit can sometimes be divided into a lower gravelly sand and an upper well sorted fine sand.

Tectonics

The units of the lower part of the cliff are affected by tectonics. Only fragmentary parts of the tectonic structures are accessible, as the lower part of the cliff is mainly covered by landslide masses. The deformation consists of overturned folds. The structures are, however, cut by erosion, and mostly only the synformal part is preserved. Overthrusts intersect the plastic deformation. The observed structures are described from the east to the west. The measurements of the deformation structures are summarized in Fig. 27.

A syncline can be seen in the stratified sediments at 7:0-10. The eastern limb has rotated 132° and is probably the lower limb of an overturned fold. The fold axis is in N50°W. The diamicton at 7:25 has been bent over the sand. The fold axis trends N50°W. This is probably a part of the same synclinal fold as at 7:0-10 with the fold axis oblique to the cliff wall.

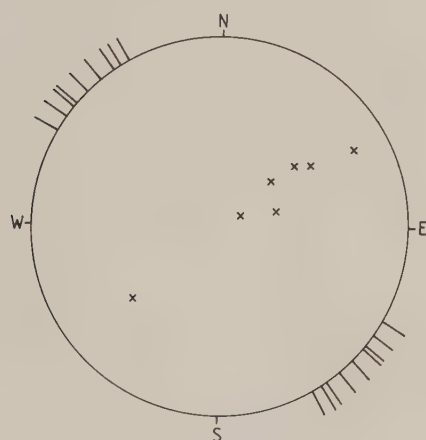


Fig. 27. Tectonics, site 7. Stereographic projection. Bedding is plotted as poles to bedding surfaces on lower hemisphere. The trend of fold axes is shown by marks outside the circle.

The overthrust at 7:80-90 has a N45°W strike and dips 28° towards NE. The sand in the footwall is underlain by diamicton facies, which can be seen at 7:100-110. This diamicton is involved in the deformation at 7:110-120. The deformation consists of a combination of a fold and an overthrust. A thrust-plane has been formed in the diamicton and dips towards the north-east with a N34°W strike. The strata are bent in a shallow syncline with an axis in N48°W to the east of the thrust-plane. To the west the laminae are overturned and dip towards the north-east with a N50°W strike.

Another overthrust can be seen at 7:160-170. The diamicton has been sheared over the sand. A crude foliation has been formed in the sand parallel to the thrust-plane.

The synclinal part of an overturned fold can be seen in the sand at 7:200-205. The fold axis trends N35°W.

5.1.8 Site 8

The north-west cape of the island is called Västernäs. The cliff is 8-11 m high, and the 280 m long section (Fig. 30) has a trend progressively changing from S70°W in the east to S30°W in the west. Several of the stratigraphic units were traced along the entire section. The stratigraphic recording and the gravel analyses were made in two portions, one from 8:0-140 (Fig. 31) and the other from 8:140-280 (Fig. 32).

Lithostratigraphic units, section 8:0-140

Unit 8a:1(Facies Elf and Elp). The unit can be traced as a 1.5 m thick unit in the basal part of the cliff. The lower boundary lies below the sea level. A fissile facies can be studied in the eastern part of the section, but the main part of the unit consists of a structureless, homogeneous diamicton with subhorizontal widely spaced sand partings. The silt and clay content is around or slightly less than 50%. Striated boulders are quite frequent. Smoothed stoss-sides and rough lee-sides were generally recognized. Striae directions were from N48°E to S68°E (Fig. 28). Fabric analyses were made at 8:35 and 8:115. The rose diagrams have main peaks at N40°W/S40°E and N30°W/S30°E.

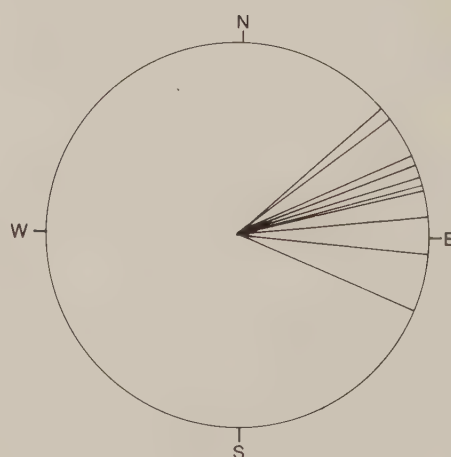


Fig. 28. Striae directions, measured on boulders in unit 8a:1. The directions are decided by the orientation of stoss- and lee-sides on the boulders.

Site 7, section 7:0-240



Fig. 29. Section chart, site 7.



Site 8, section 8:0-140



section 8:140-280

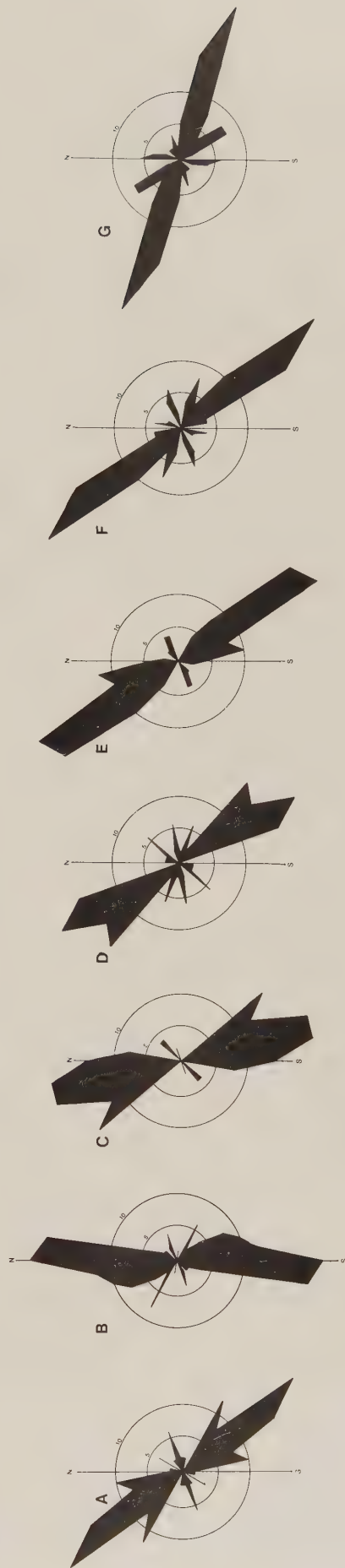
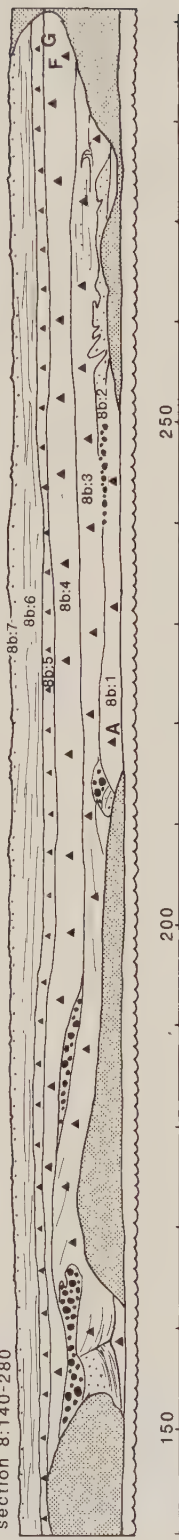


Fig.30. Section chart, site 8. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

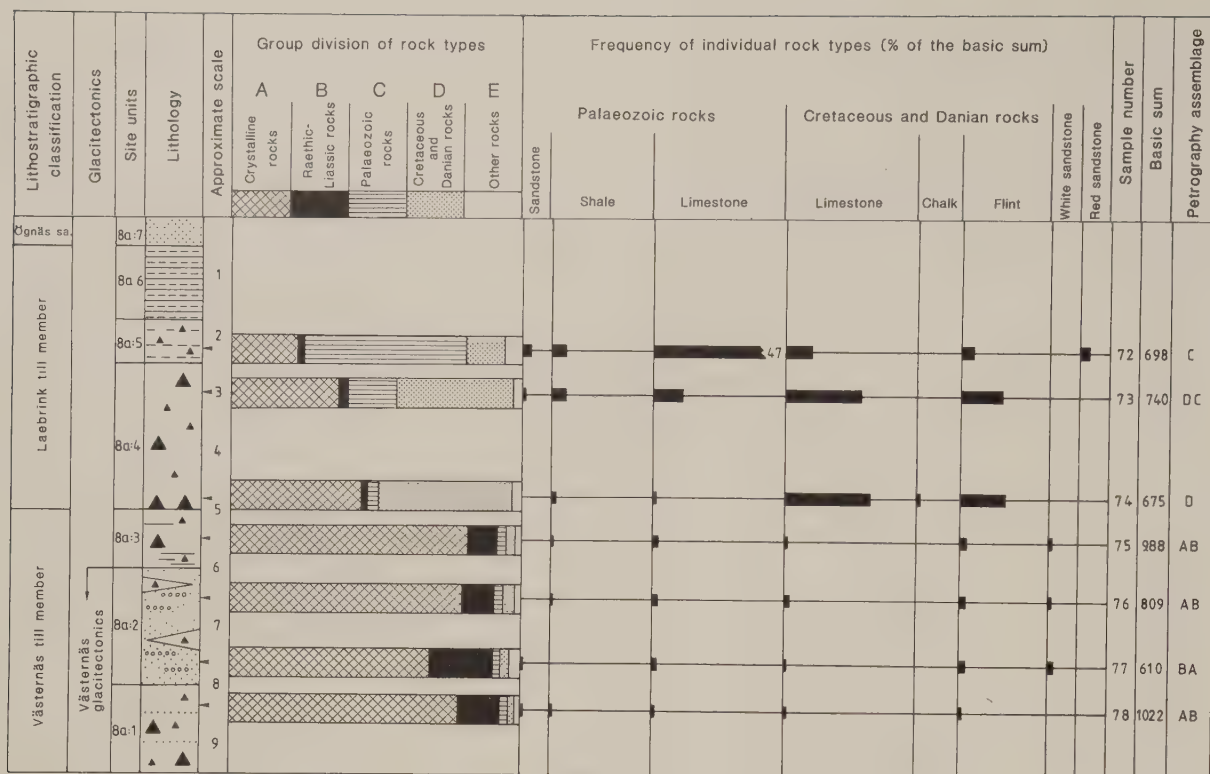


Fig. 31. Lithostratigraphy and petrography spectra, site 8, part 8:0-140.

Unit 8a:2(Facies B and E3B). A heterogeneous unit was traced from 8:70 and further westwards. The unit is dominated by sand facies but the sand is interfingered by lenses of diamicton and gravel. The ratio of diamicton beds increases westwards. The unit is about 2 m thick.

Unit 8a:3(Facies E2). This unit extends from 8:80 to 8:125. It lies discordant on the deformed lower units. The diamicton is crudely stratified or banded and has a high content of sorted sand. Two boulders, buried in the substratum at 8:95 and 8:108 and with their upper surfaces in contact to the 8a:3 unit, were striated on a bevelled surface. The striae were from N50°E and N55-60°S.

Unit 8a:4(Facies E1f). The diamicton lies with a sharp boundary to the lower units. The thickness of the unit decreases from 3.1 m in the east to 0.2 m at 8:120. The diamicton is homogeneous and has a silt and clay content of approximately 50%. A boulder enrichment occurs in the middle part of the unit between 8:0 and 8:80. A slight boulder enrichment was also noticed at the lower boundary. The boulders were partly buried in the substratum. A series of fabric analyses was made at 8:35 from the base to the top of the unit. The vertical distance between the sets of reading was 0.5 m. The preferred orientation of clasts, from the base to the top, is N5°E/S5°W, N20°W/S20°E, N15°W/S15°E, N35°W/S35°E and N35°W/S35°E.

Unit 8a:5(Facies E3D). The fissile diamicton is overlaid by 0.7-1 m of stratified diamicton. Scattered observations in the upper part of the vertical wall were made from a ladder. The unit is similar to the 7:6 unit, with thin diamicton beds and intrabeds of silt and clay.

Unit 8a:6(Facies D). A homogeneous or crudely stratified clay with a thickness of 0.75 to 1.5 m lies on top of the diamicton. The clay contains outside clasts.

Unit 8a:7(Facies B). The uppermost 0.4-0.5 m consist of sand with a minor content of gravel.

Lithostratigraphic units, section 8:140-280

Unit 8b:1(Facies E1p). The diamicton can be traced in the lower 1.5 m along the entire section. It continues below the beach level. The diamicton is homogeneous but separated into 0.1 to 0.2 m thick subhorizontal units by sand partings. Two fabric analyses were made at 8:220 in the lower and upper part of the unit. Both the diagrams have their main peaks at N30°-40°W/S30°-40°E.

Unit 8b:2(Facies A and B). A discontinuous unit of gravel and sand lies on top of the diamicton. The gravel dominates in the eastern part between 8:140 and 8:200 and sand predominates at 8:230-260. The sediments are folded and thrust.

Unit 8b:3(Facies E1-2). This unit is very similar to unit 8b:1. Grain size composition, orientation of clasts and sand partings are the same. The upper part of unit 8b:3 is, however, more sandy and more heterogeneous in texture.

Unit 8b:4(Facies E1f). The unit is the same as unit 8a:4. Upper and lower boundaries are sharp, and the average thickness is 2-3 m. The preferred orientation of clasts in the middle of the unit at 8:210 is N35-45°W/S35-40°E. Two analyses at 8:270 gave preferred orientations at N40°W/S40°E at 1 m below the upper boundary and N70-80°W/S70-80°E at 0.2 m below the upper boundary.

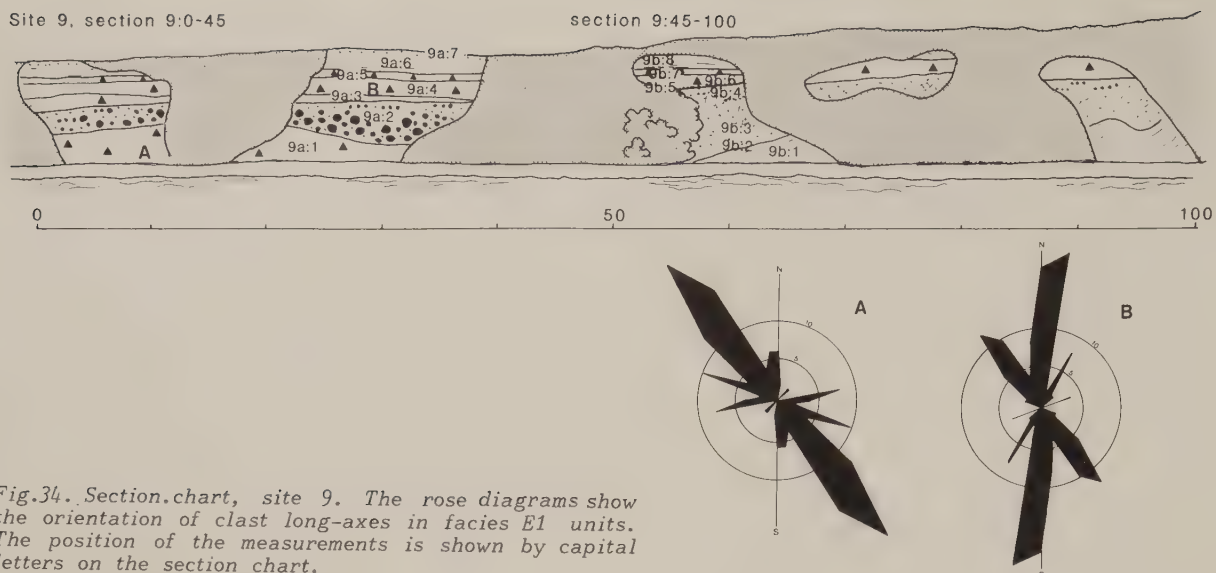


Fig.34. Section chart, site 9. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

9:45-100 (Fig. 36). The general trend of the cliff is S10°E, and the top of the cliff rises from 11 m a.s.l. in the north to 15 m a.s.l. in the south.

Lithostratigraphic units, section 9:0-25

Unit 9a:1 (Facies E1p). The base of the cliff consists of a homogeneous diamicton with sand partings. The lower boundary of the unit was not visible in the cliff, and the exposed part of the unit is 1.2 m thick. A fabric analysis in the diamicton shows a preferred orientation of clasts at N40°W/S40°E.

Unit 9a:2 (Facies A). The contact between the diamicton and the gravel is erosive. The gravel has a maximum thickness of 4.3 m at 9:15-20 and decreases in thickness in both directions. The gravel is poorly sorted, and the content of cobbles and boulders is high. There is a fining-upward trend in the gravel, and the uppermost part consists of cross-bedded and planar laminated gravelly sand.

Unit 9a:3 (Facies E2f and E1f). The gravelly sand at the boundary to the diamicton is penetratively sheared, and sand and diamicton are mixed in a banded zone in the lower part of the diamicton unit. The sand lenses or bands decrease upwards, and the diamicton becomes more homogeneous in the upper part. The diamicton is not continuous along the section and has a maximum thickness of 1.5 m.

Unit 9a:4 (Facies E1f). The contact between the two diamictons is sharp. The 9a:4 unit has a thickness of approximately 2.5 m. One fabric analysis was made in the lower part of the unit, and shows a preferred orientation of clast longaxes at N10°E-S10°W. A boulder, buried in the 9a:3 unit and with the upper surface at the lower boundary of the 9a:4 unit was striated from S 4°W.

Unit 9a:5 (Facies E3D). 0.3-0.8 m of distinctly stratified diamicton overlayers the fissile diamicton. The individual diamicton layers are thin, seldom exceeding 10 m. Palaeozoic limestone clasts are very frequent and the diamicton layers have a characteristic greenish gray colour. The silt and clay intrabeds generally lie undisturbed in horizontal la-

minae. Loadclasts and erosive contacts are present but rare.

Unit 9a:6 (Facies D). The diamicton is covered by 0.8-1.0 m of clay. The lower part has dispersed clasts, mostly of palaeozoic limestone.

Unit 9a:7 (Facies B). The top of the sequence consists of fine to medium sand. Diffuse horizontal zones with a high humus content are present. Gravelly sand was sometimes found in the lower part of the sand unit.

Lithostratigraphic units, section 9:45-100

Unit 9b:1 (Facies D). The basal part of the cliff at 9:60 and 9:100 consists of laminated silt and clay. Thick beds of homogeneous and convolute silt occur in the stratified sediments, and a few outside clasts are embedded in the laminated parts. The thickness of the exposed strata is 3.5 m, but the unit continues below the present sea level.

Unit 9b:2 (Facies C). There is a gradual change from alternating silt and clay to alternating sets of cross-laminated sand and clay laminae. The thickness of individual sand beds is 10 to 25 cm, and the total thickness of alternating bed facies is about 1 m.

Unit 9b:3 (Facies B). The alternating bed facies alters into 4.5 m of sand. Cross-lamination dominates in the lower part and planar lamination in the upper part.

Unit 9b:4 (Facies A). Fine gravel becomes gradually more abundant, and the upper 1 m of the stratified sediments consists mainly of planar laminated and cross-bedded sandy gravel and gravelly sand. The whole coarsening-upward sequence, unit 9b:1 to 9b:4, is affected by tectonism and is intersected by numerous reverse faults.

Unit 9b:5 (Facies A). The dislocated strata are truncated by a thin, discontinuous bed of gravel. The maximum thickness is 20 cm.

Unit 9b:6 (Facies E1f). The unit was traced into the 9a:4 unit.

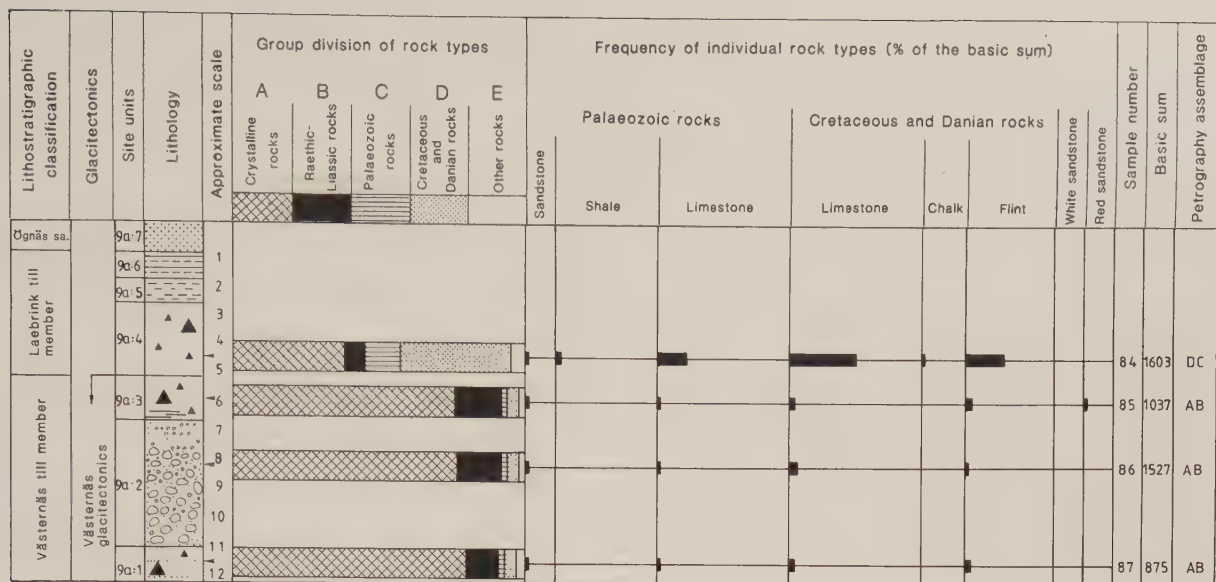


Fig. 35. Lithostratigraphy and petrography spectra, site 9, part 9:0-30.

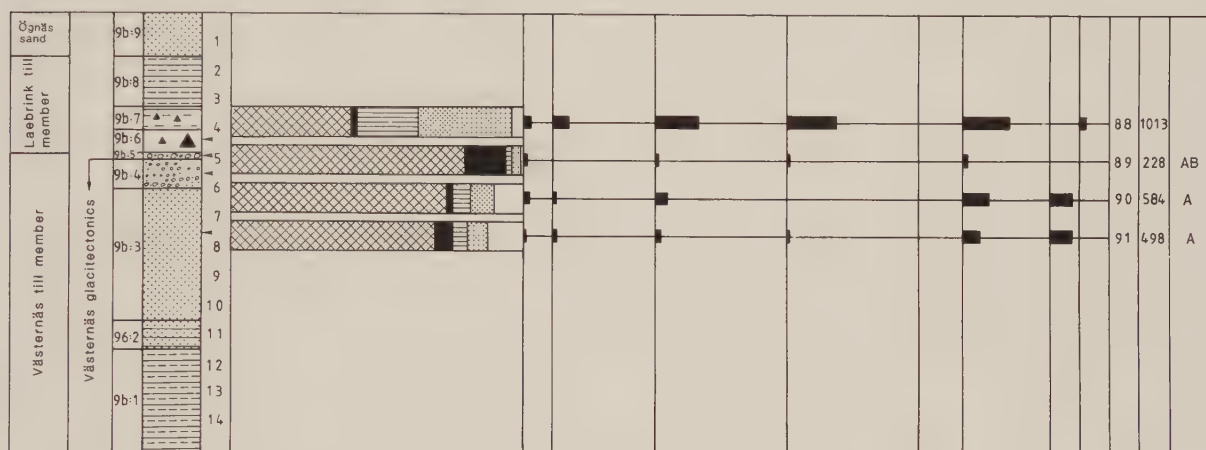


Fig. 36. Lithostratigraphy and petrography spectra, site 9, part 9:45-100.

Unit 9b:7(Facies E3D). The unit is equal to unit 9a:5.

Unit 9b:8(Facies D). The unit is equal to unit 9a:6.

Unit 9b:9(Facies B). The unit is equal to unit 9a:7.

Tectonics

Units 9b:1 to 9b:4 are affected by glacitectionism. The general dip of bedding planes in the lowermost part is towards NE. The sand and gravel in the upper part are intersected by a mosaic of minor faults. The tectonic structures were not measured.

5.1.10 Site 10

Site 10 comprises the 17 to 20 m high cliff to the north of the Kyrkbacken village. The section is 300 m long and trends S10°E (Fig. 37). The cliff section has two main outcrops, one from 10:0 to 10:120, where the lower and the middle part of the strata are exposed, and

one from 10:200 to 10:300. The lithology and the petrography composition of the strata is shown in Fig. 38 (10:0-120) and Fig. 39 (10:200-300).

Lithostratigraphic units, section 10:0-120

Unit 10a:1(Facies B and C). The main part of the cliff consists of sand. The grain size composition and the sedimentary structures change^a from north to south. The northern part contains cross-bedded and planar laminated sand with a minor content of fine gravel. The middle part is dominated by cross-laminated and planar laminated sand, and the southern part of the sand has a minor content of laminated silt and clay. The sand is dislocated.

Unit 10a:2(Facies E2f-E1f). The diamicton lies with an angular unconformity on the dislocated lower strata. The basal part of the diamicton is banded and consists mainly of crudely laminated sand. The contact to the stratified sand is very sharp. The crudely laminated or banded lower part of the diamicton grades into a more homogeneous fissile

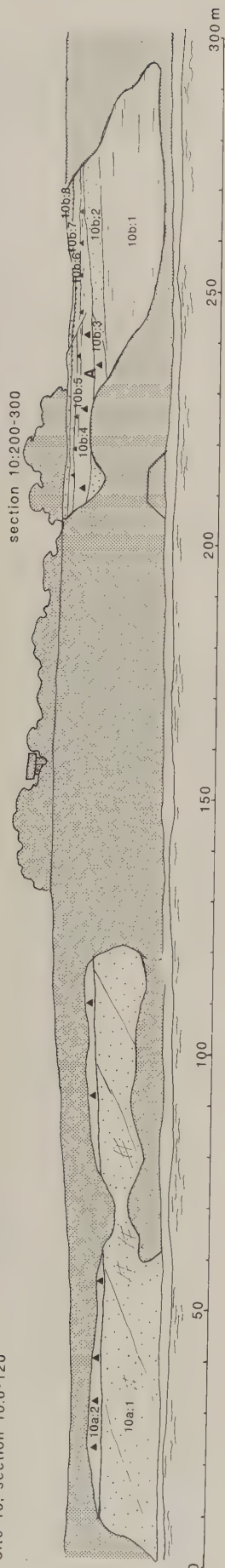


Fig. 37. Section chart, site 10. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

diamicton. The upper part of the cliff is covered. A clay unit and a 2.5 m thick unit were, however, traced on top of the diamicton.

Lithostratigraphic units, section 10:200-300

Unit 10b:1(Facies D). The lower part of the cliff at 10:200-300 consists of silt and clay. The unit continues below the present sea level. The silt and clay is mainly laminated but contains some thick beds of homogeneous and convoluted silt. Slump scars are sometimes connected to the thicker beds. The grain size composition increases slightly upwards, and some thick beds 0.4 to 0.75 m of coarse silt and fine sand with convolute and dish structures are present in the upper half of the 12.5 m thick unit. The strata of unit 10b:1 are slightly tectonized but the bedding is almost subhorizontal.

Unit 10b:2(Facies B). The boundary between the silt and clay facies and the sand is an almost parallel unconformity with indication of erosion. The surface is overlaid by 3 m of planar laminated sand at 10:250-275.

Unit 10b:3(Facies E1f). The unit is laterally restricted to a small depression between 10:230 and 10:250. The unit has a maximum thickness of 2.5 m at 10:240.

Unit 10b:4(Facies E1f). The 10b:4 diamicton lies with a sharp boundary to the lower diamicton and continues to the north as a 3 m thick bed. The thickness decreases to 0.5 m in the southern part of the section. The diamicton is generally homogeneous. The lowest part has a minor content of sand and silt banding. A fabric analysis at 10:235 shows a preferred orientation of clast longaxes at N10°W/S10°E. The reading was made 10 to 20 cm from the base of the unit.

Unit 10b:5(Facies E3C and C). The diamicton is capped by stratified silt and sand, which also contain thin layers of diamicton. A few isolated cobbles were found in the stratified sediments. The silt laminae are bent around the clasts. In some cases a small-scale folding was found below the clasts. The thickness of the unit varies from 0.2 to 1.0 m.

Unit 10b:6(Facies D). The sequence is succeeded by 1 m of faintly laminated or almost homogeneous silty clay at 10:230. The thickness decreases towards the south.

Unit 10b:7(Facies A). A very thin layer of gravel, or rather a gravel lag lies on top of the clay.

Unit 10b:8(Facies B). The top of the sequence is a 1 to 1.5 m thick fine sand unit. The sand is structureless but contains diffuse layers of humic sand.

Tectonics

The sand at 10:0-120 is intersected by numerous low-angle large thrust faults. The vertical displacement is small, generally less than 1 m. High angle fault systems of second order are often seen in the foot wall (Fig. 40). The direction of first order thrust planes and the general dip of bedding planes in the thrust blocks are displayed in Fig. 41. A very shallow system of faults can be seen just below the sole of the 10a:2 diamicton in connection with a large boulder at 10:40.

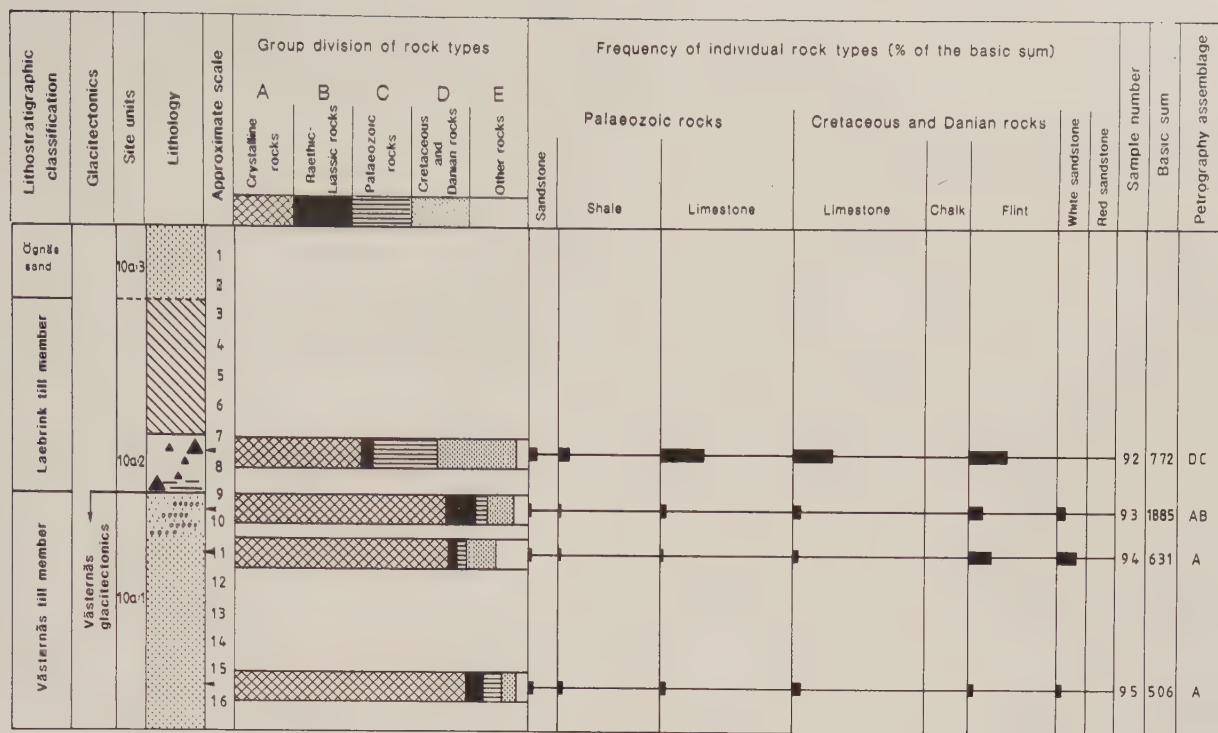


Fig. 38. Lithostratigraphy and petrography spectra, site 10, part 10:0-120.

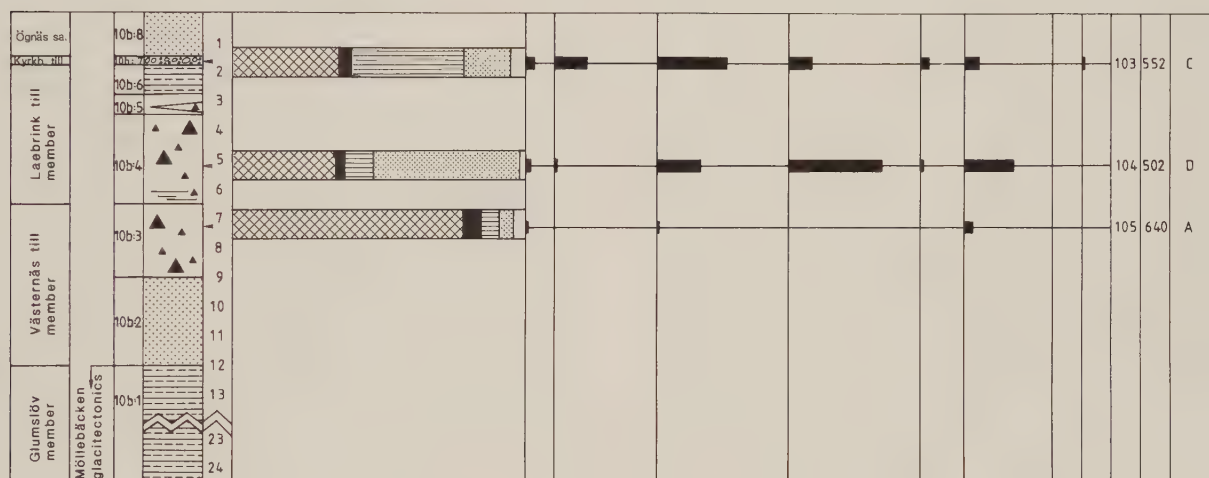


Fig. 39. Lithostratigraphy and petrography spectra, site 10, part 10:200-300.

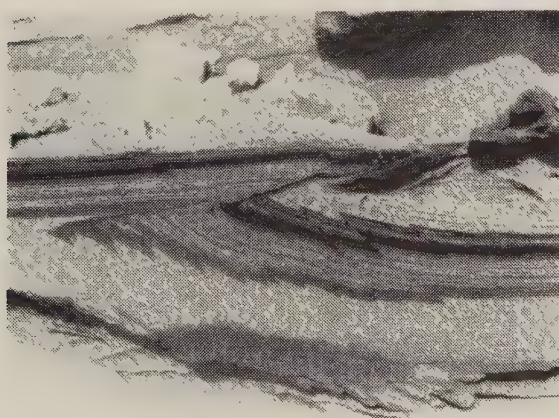


Fig. 40. First and second order fault systems in thrust sand of unit 10a:1.

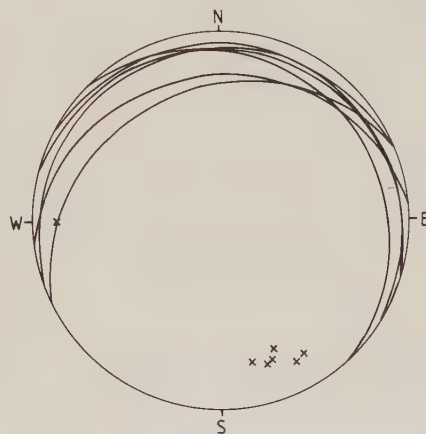
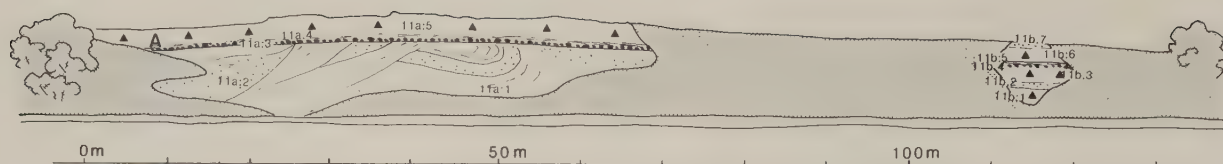


Fig. 41. Tectonics, site 10. Stereographic projection.



5.1.11 Site 11

The strata in the steep coast to the north of the Ålabodarna fishing-village are mainly covered by slumps and vegetation. In some parts the erosion is more intense and has resulted in almost vertical cliff cuttings. The largest outcrop is situated in the highest part of the cliff about 1 km to the north of the fishing harbour. The top of the cliff lies 17 m above the sea level and declines towards the south in the section. The section (Fig. 42) trends S35°E. The stratigraphy of site 11 is divided into two sequences, one for the main outcrop (Fig. 43), and one for the minor outcrop 50 m further to the south (Fig. 44).

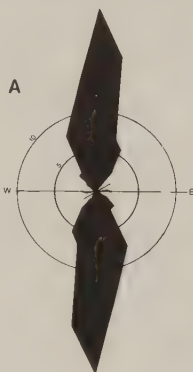


Fig. 42. Section chart, site 11. The rose diagram shows the orientation of the clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

Lithostratigraphic units, section 11:0-65

Unit 11a:1(Facies D and B). The main part of the outcrop consists of laminated silt and clay and thicker beds of convoluted and homogeneous silt. A 2 m thick bedset of cross-laminated very fine sand and coarse silt can be seen in the upper middle part of the unit. The upper boundary of the unit is a discordant erosive contact to younger sediments. The lower boundary was not seen in the cliff.

Unit 11a:2(Facies B). The lower part of the cliff at 11:0-15 consists of sand with a minor content of fine gravel. Cross-bedding and planar bedding are the most common structures. The stratigraphic relationship between unit 11a:1 and unit 11a:2 can be interpreted from the boundary of angular unconformity. The bedding in the sand is parallel to the contact surface, while the boundary truncates the bedding of the silt and clay. The contact surface dips 35° towards north-west, and neither the silt and clay, nor the sand, lie in a primary position.

Unit 11a:3(Facies A). A thin continuous bed of gravel, 0.2 to 0.7 m thick, truncates the bedding of the two lower units. The gravel is poorly sorted and contains several cobbles and boulders. Lenses of diamicton and laminated sand are abundant.

Unit 11a:4(Facies D). The gravel is covered by 0.1 to 0.2 m of faintly laminated silt and clay.

Unit 11a:5(Facies Elf). The top unit in the cliff is a fissile diamicton, which ranges in thickness from 2.8 m to 3.7 m. One fabric analysis, performed 1 m from the base of the unit at 11:10, gave a preferred orientation of clast longaxes at N5°E/S5°W.

Unit 11b:2(Facies B). A thin unit, 5 to 10 cm thick, of gravelly sand lies with a sharp, slightly erosive contact to the diamicton.

Unit 11b:3(Facies E1 and E3B). A 1.8 m thick sequence of heterogeneous diamicton lies on top of the sand. The lower part of the diamicton is clayey and rather homogeneous, while the upper part consists of thin sandy diamicton layers interbedded by sand and gravel. The excavation was only 1 m wide, and the unit could not be traced into the large exposure to the north.

Unit 11b:4(Facies A). The unit is the same as unit 11a:3.

Unit 11b:5(Facies C). The unit consists of 0.9 m of cross-laminated fine sand with silt and clay laminae. It was traced into unit 11a:4.

Unit 11b:6(Facies Elf). The unit is the same as unit 11a:5. One fabric analysis in the lower part of the 1.8 m thick diamicton bed shows a preferred orientation of clast longaxes at N/S.

Unit 11b:7(Facies B). A 0.5 m thick unit of sorted sand was distinguished at the top of the sequence.

Tectonics

The most dominating tectonic feature is a synformal bend at 11:50-60. The southern limb is overturned towards the north-west. The axis of the fold is in N53°E/S53°W. The sand at 11:0-15 is transversely by some high angle reverse faults. The strikes of two of the fault planes are N45°E and N53°E. The dip of the bedding planes is 35° towards N47°W. The tectonics is shown in Fig. 45.

Lithostratigraphic units, section 11:115-117

Unit 11b:1(Facies Elf). The lower part of the small excavation consists of a silty diamicton with an extremely low content of gravel-sized and larger clasts. The rose diagram from the fabric analysis in the diamicton has a main peak at N/S.

5.1.12 Site 12

The section (Fig. 46) is situated in the same cliff and 175 m to the south of section 11. The top of the cliff lies 10 to 11 m above the sea level, and the total section is about 120 m long. The observations were made in three separate cuttings. The two northern

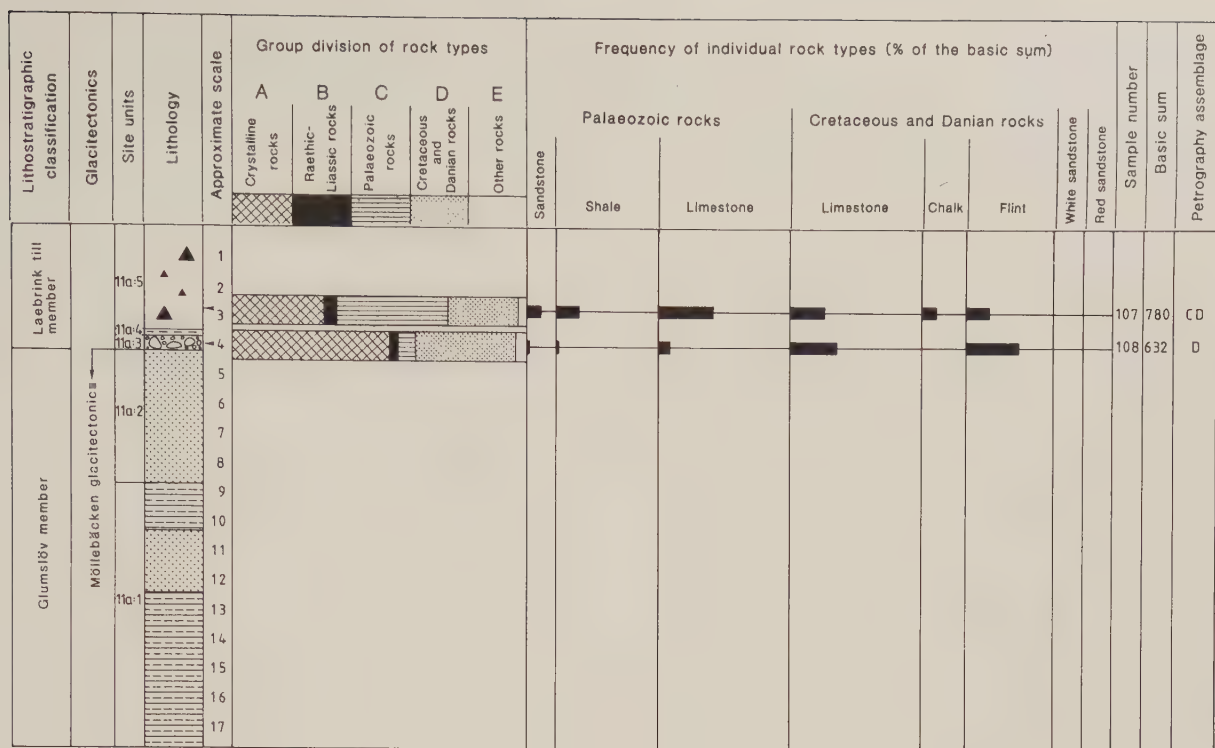


Fig. 43. Lithostratigraphy and petrography spectra, site 11, part 11:0-65.

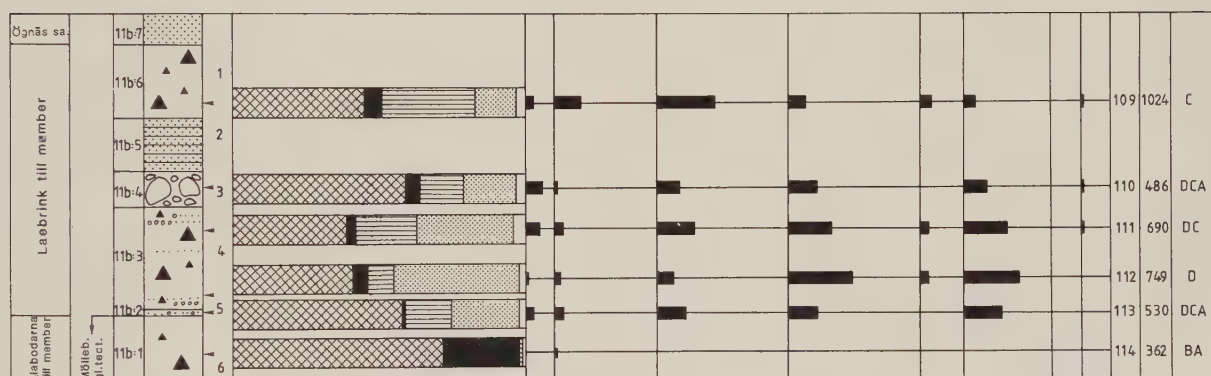


Fig. 44. Lithostratigraphy and petrography spectra, site 11, part 11:115.

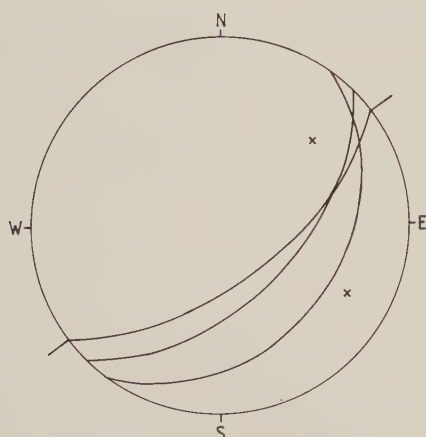


Fig. 45. Tectonics, sites 11 and 12. Stereographic projection. Thrust faults are plotted as large circles. Bedding is plotted as poles to bedding surfaces on lower hemisphere. The trend of fold axes is shown by marks outside the circle.

cuttings have a similar stratigraphy, but the southern cutting diverges in the lower part. The stratigraphy is therefore presented in two records, Fig. 47 and Fig. 48.

Lithostratigraphic units, section 12:0-70

Unit 12a:1(Facies D). This unit constitutes the lower portion of the cliff and continues below the sea level. The silt and clay is mainly laminated, sometimes with a convolute lamination. Homogeneous layers of silt also occur.

Unit 12a:2(Facies B). The contact to the lower unit is sharp, and the surface is scoured. The unconformity is parallel. Planar lamination and cross-bedding are the dominating structures in the sand. The maximum thickness of the unit, measured at 12:10, is 2.6 m.

Unit 12a:3(Facies A). The gravel lies discordantly on the sand unit. The gravel is poorly

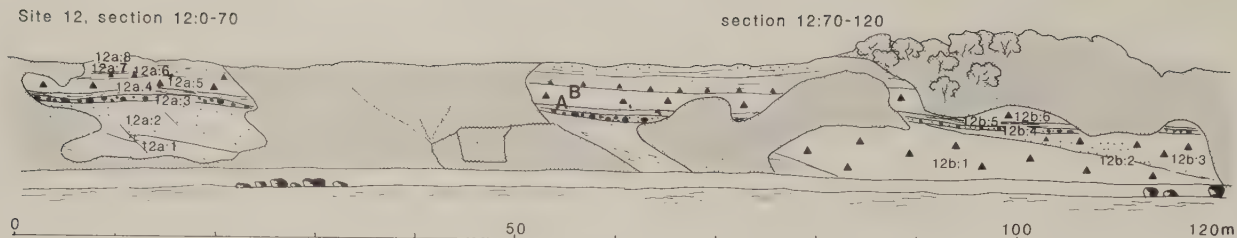
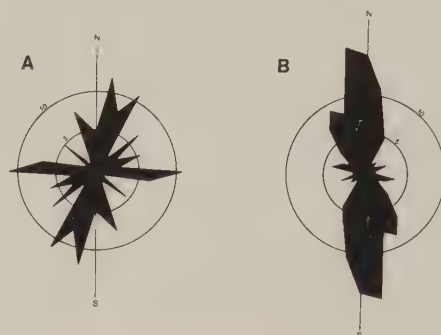


Fig.46. Section chart, site 11. The rose diagram shows the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.



sorted and has a high content of pebbles and boulders. It also contains lenses of diamicton and sorted sand. The thickness ranges between 0.2 and 0.6 m.

Unit 12a:4(Facies D). 0.2 to 0.3 m of laminated silt and clay lies on top of the gravel.

Unit 12a:5(Facies E1f). Upper and lower contacts are sharp. The thickness ranges between 2.5 and 3 m. Two fabric analyses were performed. The preferred orientation of clast long-axes in the lower part of the unit is N20°E/S20°W, and N/S in the upper part.

Unit 12a:6(Facies E3D). Approximately 1.5 m of stratified diamiction was traced on top of the fissile diamicton in the two outcrops. The diamicton layers range in thickness from

a few centimetres to 10–15 cm. Grain size composition and the petrography composition differ between the diamicton beds.

Unit 12a:7(Facies D). The unit consists of an almost structureless clay with dispersed out-size clasts.

Unit 12a:8(Facies B). The uppermost part of the cliff consists of sand.

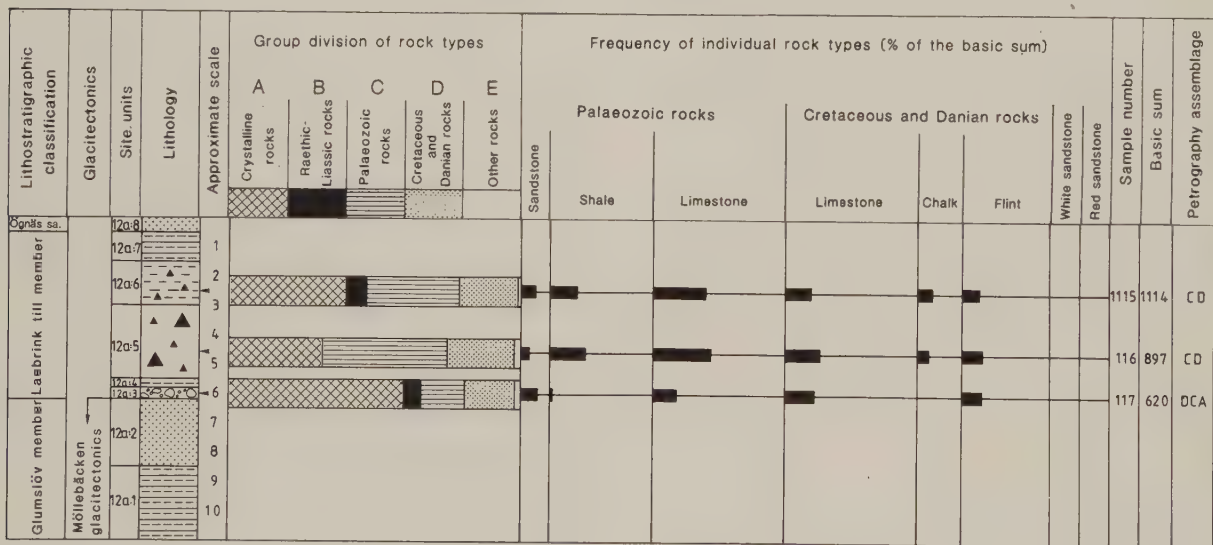


Fig. 47. Lithostratigraphy and petrography spectra, site 12, part 12:0-20.

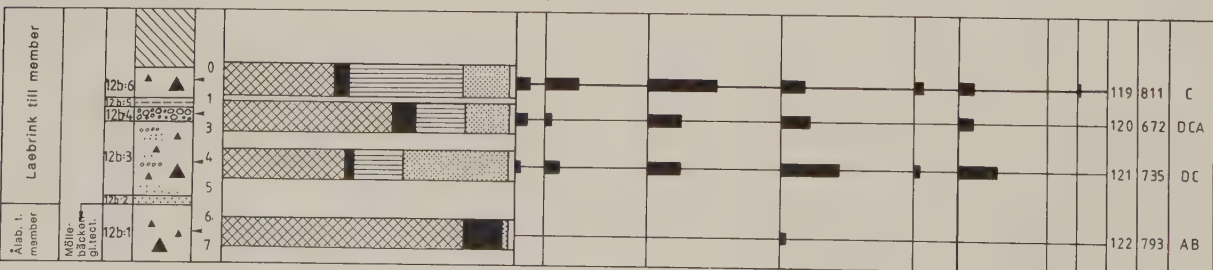


Fig. 48. Lithostratigraphy and petrography spectra, site 12, part 12:70-120.

Lithostratigraphic units, section 12:70-120

Unit 12b:1(Facies Elf). The unit is fissile, silty diamicton with very few large clasts. Elongated clasts have a preferred orientation at N/S.

Unit 12b:2(Facies B). The upper surface of the diamicton is slightly scoured and overlaid by a 10 to 25 cm thick unit of well sorted sand.

Unit 12b:3(Facies E3B). This is a wedgeformed unit, which starts at 12:75 and increases in thickness to 2.5 m at 12:120. Thick irregular or thin layers of diamicton are interstratified by sand and gravel.

Unit 12b:4(Facies A). The unit is equal to unit 12a:3. Layers of sorted sand and gravel are more dominating in this part.

Unit 12b:5(Facies D). The unit is equal to unit 12a:3. Seven distinct clay laminae, separated by laminated silt, were recognized.

Unit 12b:6(Facies Elf). Only the lower 0.5m of the unit were exposed. The contact to the silt and clay is sharp. The unit can be traced into the 12a:5 unit.

Tectonics

The dissimilarity in strata of the lower northern part in contrast to the lower southern part of the section is due to a major fault at 12:65-75. The strike of the fault is N35°E, and the fault plane dips 35° towards south-east. The hanging wall consists of the 12b:1 unit, and the 12a:1 and 12a:2 units constitute the footwall. Drag flexures have been formed on the flanks of the fault plane, and they indicate a reverse displacement. The silt and sand at 12:0-20 is intersected by numerous minor faults, none of which were measured. Measurements of tectonic structures are displayed in Fig. 45.

5.1.13 Site 13

Strandnäs is the northernmost of the two clay-pits in Ålabodarna. The pit can be divided into two parts, separated by a large almost vertical fault. The strata to the north of the fault consist of grey diamicton beds, interlayered by silt and clay. The southern part consists mainly of stratified sand. The clayey diamicton was mined from a horizontal plane in the middle of the pit. To the west of this plane the clay was brought up from a deep ditch, and to the east the sediments were taken from the pit wall. The ditch is now infilled, the slopes are covered by vegetation, and the information was gathered from minor outcrops and excavations in different parts of the pit. The lithostratigraphy is shown in Fig. 49.

Lithostratigraphic units, site 13

Unit 13:1(Facies E1 and E3D). The measured thickness of this unit is 21 m. The lower boundary was, however, not reached in the bottom of the pit, and the figure is thus only a minimum value. A minor outcrop in the western side, close to the bottom of the clay pit, shows an almost homogeneous diamicton with very few and scattered gravel-sized or larger clasts. A fabric analysis in this

homogeneous part gave a preferred orientation of clast longaxes at N5°E/S5°W. The diamicton was also studied at minor outcrops in the northern part of the pit and in a major cutting in the eastern wall. Those outcrops are situated at a higher level in the pit and represent a higher stratigraphic level in the unit. In these outcrops the diamicton is stratified and interlayered by silt and clay. The individual diamicton layers are faintly laminated. The lamination is seen as colour variations in different shades of grey. Some layers have intraclasts of clay.

Unit 13:2(Facies B). The upper surface of the diamicton in the eastern wall is scoured and overlaid by a thin layer of gravel and by sand. The sand also outcropped in the former ditch in the southern part of the pit and can be reached only by post auger drillings from the plane at the toe of the south-eastern pit wall. The thickness of the sand unit, measured in such a drilling, was 4.3 m. Fossile ice wedge polygons in the sand have been studied during excavations 1962 by Johnsson (1984).

Unit 13:3(Facies C). The facies changes to alternating bed facies with distinct clay laminae, interbedded by cross-laminated sand.

Unit 13:4(Facies D). The sand between the clay laminae is replaced by parallel-laminated and cross-laminated silt. The unit also contains more homogeneous silt layers and layers with convolute lamination. This unit was excavated in the lower part of the south-eastern pit wall and has a thickness of 2.5 m.

Unit 13:5(Facies C). The distinct clay laminae remain the same, but the sediments between the clay laminae become sandier. The vertical distance between the clay laminae also becomes successively wider, and the clay laminae finally cease 5 m higher up in the wall.

Unit 13:6(Facies B). The uppermost part of the faulted sediments in the southern part of the pit consists of sand with a minor content of silt. Fossile ice-wedge casts have been observed by Johnsson (1962,1963) and by the author. The measured thickness of the sand unit is approximately 7.5 m, but the upper part is cut off by erosion. The sequence shows a slight coarsening upwards with trough cross-bedded and planar-laminated gravelly sand in the uppermost part.

Unit 13:7(Facies Elf). The dislocated lower strata, units 13:1 to 13:6, are cut off by an angular unconformity and are overlaid by a fissile diamicton. The preferred orientation of clast longaxes is N25°W/S25°E. A larger fossile ice wedge, infilled by diamicton from unit 13:7, penetrates down into the sand substratum (unit 13:2) in the northern part of the pit wall. The upper part of the diamicton has been removed by a bulldozer, and only 1 m was left in the wall. The strata that once covered the diamicton are unknown.

Tectonics

The large fault, which divides the strata in the pit into two parts, is evident by the spatial distribution of the strata. The vertical displacement can also be estimated to 15-20 m. The fault itself was traced a short distance in the eastern wall. It consists of two parallel faults, 15 cm apart, with a N60°W strike and a dip of 84° towards the

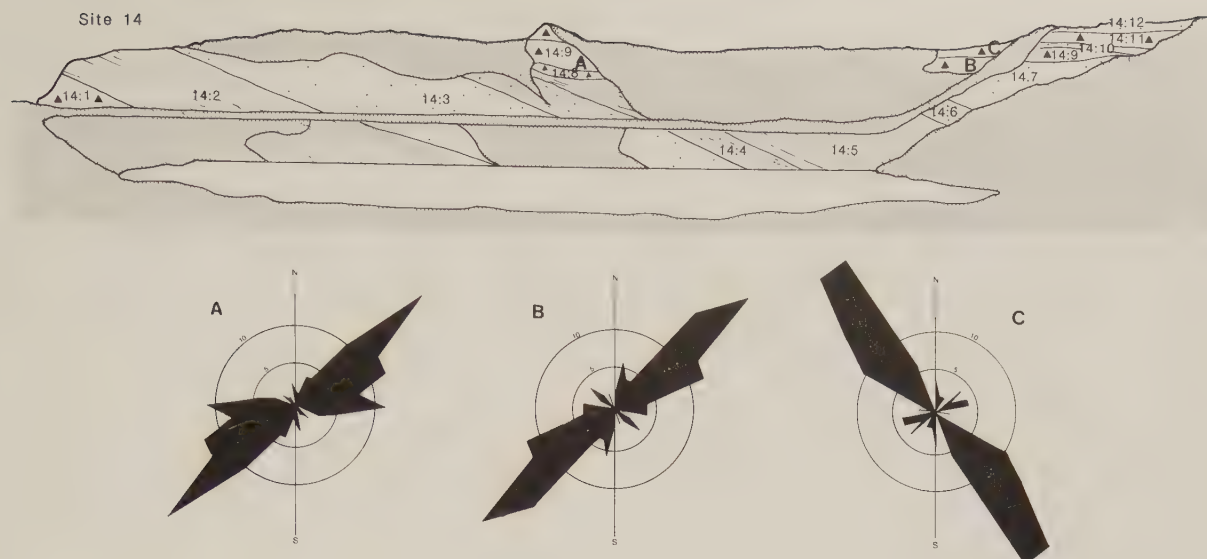


Fig. 51. Section chart, site 14. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

were made in the north-west part of the pit. As in the Strandnäs pit, the excavation was made in a wall and in a deep ditch, separated by a horizontal plane. The lower strata are dislocated and tilt towards the north-east. Lithostratigraphy and gravel analysis results are shown in Fig. 52.

Lithostratigraphic units, site 14

Unit 14:1(Facies E3D). The lowest unit was found in the western part of the pit. It consists of silty diamicton layers, most of them crudely laminated, interlayered by laminated silt and clay.

Unit 14:2(Facies D). The stratified diamicton is overlaid by laminated silt and clay. The clay laminae are 0.5–1.5 cm thick and separated by parallel-laminated and cross-laminated silt. Some thick beds of homogeneous silt and convolute laminated silt occur in the approximately 7 m thick sequence.

Unit 14:3(Facies B). The contact between unit 14:2 and 14:3 is distinct and shows evidence of scouring. Some shallow channels were eroded on the upper surface of the silt and clay unit. A fossil ice wedge was also found on the surface. The wedge was infilled by sand. The approximate thickness of the sand is 5 to 7 m.

Unit 14:4(Facies C). The unit consists of distinct clay laminae, interlayered by 10 to 20 cm thick sets of cross-laminated sand. The alternating bed facies is a 2 m thick transitional zone between the sand facies and the next unit, which is of silt and clay facies.

Unit 14:5(Facies D). The sand between the clay laminae is replaced by laminated silt. Thicker silt layers with massive structure or with convolute lamination occur.

Unit 14:6(Facies C). The unit is similar to the 14:4 unit with clay laminae, interlayered by cross-laminated sand. The total thickness of the three units 14:4, 14:5 and 14:6 is about 10 m.

Unit 14:7(Facies B). Another gradual transition changes the facies into sand. The structures, at least in the upper part of the unit, are dominated by cross-bedding and planar lamination. The unit is the uppermost of the dislocated tilted strata.

Unit 14:8(mainly Facies A). A high plinth was left in the north-east part of the pit during the mining. The base and the lower part of the plinth consisted of tilted beds of silt and clay facies and alternating bed facies. The tilted sediments were truncated by a complex unit of gravel, sand and thin diamicton layers. The individual layers had a restricted lateral extent. The main part of the unit was a poorly sorted gravel with cobbles and boulders and with a thickness of 2 m at the eastern edge of the plinth decreasing to zero towards the west.

Unit 14:9(Facies E2f). The 14:8 unit was covered by a diamicton with some large boulders. The boundary between the two units was indistinct. The diamicton was very heterogeneous with sand lenses and a faint sand banding of diffuse lamination. A fabric analysis gave a good preferred orientation of clast longaxes at N50°E/S50°W. The unit was also traced in the northern pit-wall, where it lay direct on the dislocated units. A fabric analysis gave the same preferred orientation as the analysis in the plinth.

Unit 14:10(Facies B). The diamicton in the northern wall was overlaid by 0.4 m of cross-laminated sand.

Unit 14:11(Facies E1f). A homogeneous fissile diamicton lay on the sand in the northern wall and on the 14:9 unit on the top of the plinth. One fabric analysis was made in the northern wall. The rose diagram has a main peak at N40°W/S40°E.

Unit 14:12(Facies B and A). The top of the sequence in the northern wall consisted of 1.3 m of sand and fine gravel. The upper surface was planed by a bulldozer, and an unknown thickness of strata was removed. As in the Strandnäs clay-pit it is uncertain whether just the topsoil was removed.

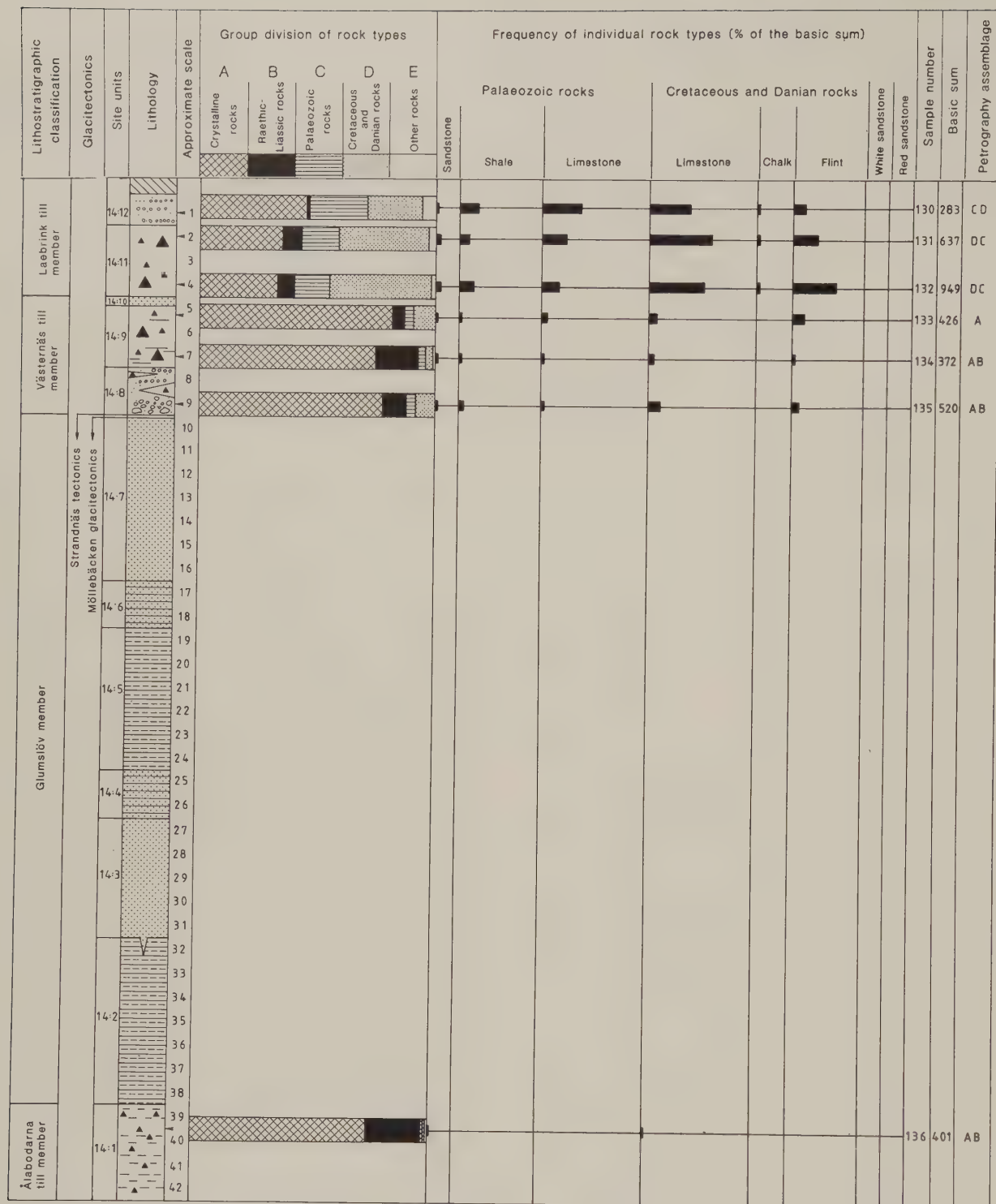


Fig. 52. Lithostratigraphy and petrography spectra, site 14.

Tectonics

The lower strata have a general dip towards the north-east. Bedding planes in unit 14:3 were measured and had a 30° dip towards N35°E. Minor faults were abundant, and two overthrusts were measured in the sand to the west of the plinth. One thrust plane had a N40°W strike and a 40° dip towards the south-west, and the other had a N32°W strike and a 40° dip towards the south-west. The measurements are shown in Fig. 53. Madsen

(1917) described a large normal fault in the Sundvik clay pit. The fault had a displacement of 11 m and a dip of 45° towards S40°W. The fault could not be rediscovered during my investigations and was probably situated further sea-wards.

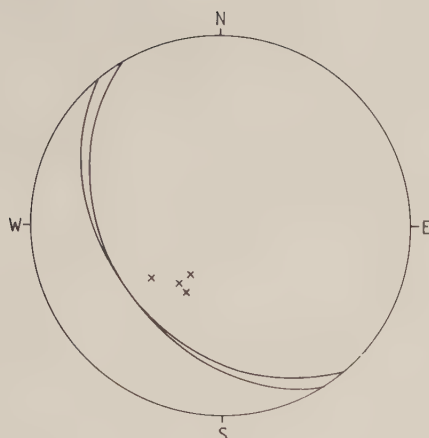


Fig. 53. Tectonics, site 14. Stereographic projection. Thrust faults are plotted as large circles. bedding is plotted as poles to bedding surfaces on lower hemisphere.

5.1.15 Site 15

Site 15 is the former clay-pit at Rustningshamn and is situated on the north side of the mouth of the Hilleshög valley. The pit walls are now sloped, but in 1973 it was still possible to excavate good sections in the north-west pit wall (Fig. 54). Some complementary studies were also made in the old pit 200 m further to the north along the coast. The lower units were affected by glaci-tectonism and dip towards the north-east. The stratigraphy is shown in Fig. 55.

Lithostratigraphic units, site 15

Unit 15:1(Facies D). The unit consists of laminated silt and clay with thick convoluted or homogeneous layers of silt. Only the upper 4 m were excavated, but the unit was traced at least 10 m further downwards in the slope.

Unit 15:2(Facies A). The boundary between unit 15:1 and unit 15:2 was sharp, and the upper surface of the silt and clay unit was scoured. The gravel covered the surface as a 1 to 2 cm thick continuous layer. Intra-clasts of silt and clay were frequent in the gravel.

Unit 15:3(Facies B). The gravel was covered by 4 m of sand.

Unit 15:4(Facies C). The sand graded into alternating bed facies with 10 to 80 cm thick sand layers, separated by clay laminae. Some of the fine sand layers were convoluted. The total thickness of the unit was about 3 m.

Unit 15:5(Facies D). The grain size composition of the layers between the clay laminae changed into silt. Only 3 m of silt and clay facies were excavated, and the upper part of the unit was covered by slumps. The unit was traced upwards in the slope by drilling.

Unit 15:6(Facies B). 5 m higher up in the slope there was a transition into sand facies. The sand was traced for at least 6 m, and minor excavations in the sand showed both cross-bedding, cross-lamination and planar lamination structures. All the units 15:1 to 15:6 dip towards the north-east.

Unit 15:7(Facies A, B and C). The tilted units were truncated by an erosive surface. The surface was slightly undulating and sloped downwards to the south-west. Unit 15:7 was restricted to the lower part in the west, and a maximum thickness of 1.45 m was measured. The unit consisted of thin beds of gravel, sand and diamicton, with the gravel as the dominating facies.

Unit 15:8(Facies Elf). The unit had a maximum thickness of 4 m in the south-west part of the outcrop, and the thickness decreased towards the north-east. The basal part, approximately the lower 0.2 m, was very heterogeneous with sand and silt banding. The lower contact, both to unit 15:7 and to the dislocated sediments, was, however, sharp. The main part of the diamicton was homogeneous with several boulders scattered in the cutting. The orientation of clast longaxes was measured 1.5 m from the base of the unit, and the preferred orientation was N35°E/S35°W.

Unit 15:9(Facies Elf). The contact between the two diamicton units was sharp. A slight concentration of boulders was seen in the boundary. The top of unit 15:9 was not reached in the outcrop, as the upper part of

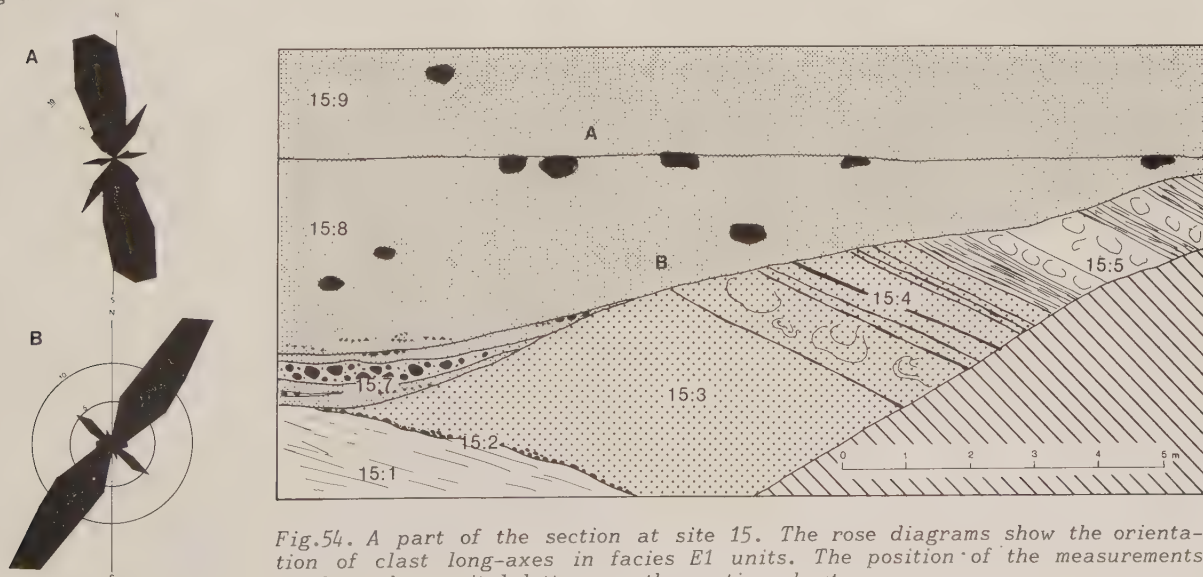


Fig. 54. A part of the section at site 15. The rose diagrams show the orientation of clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

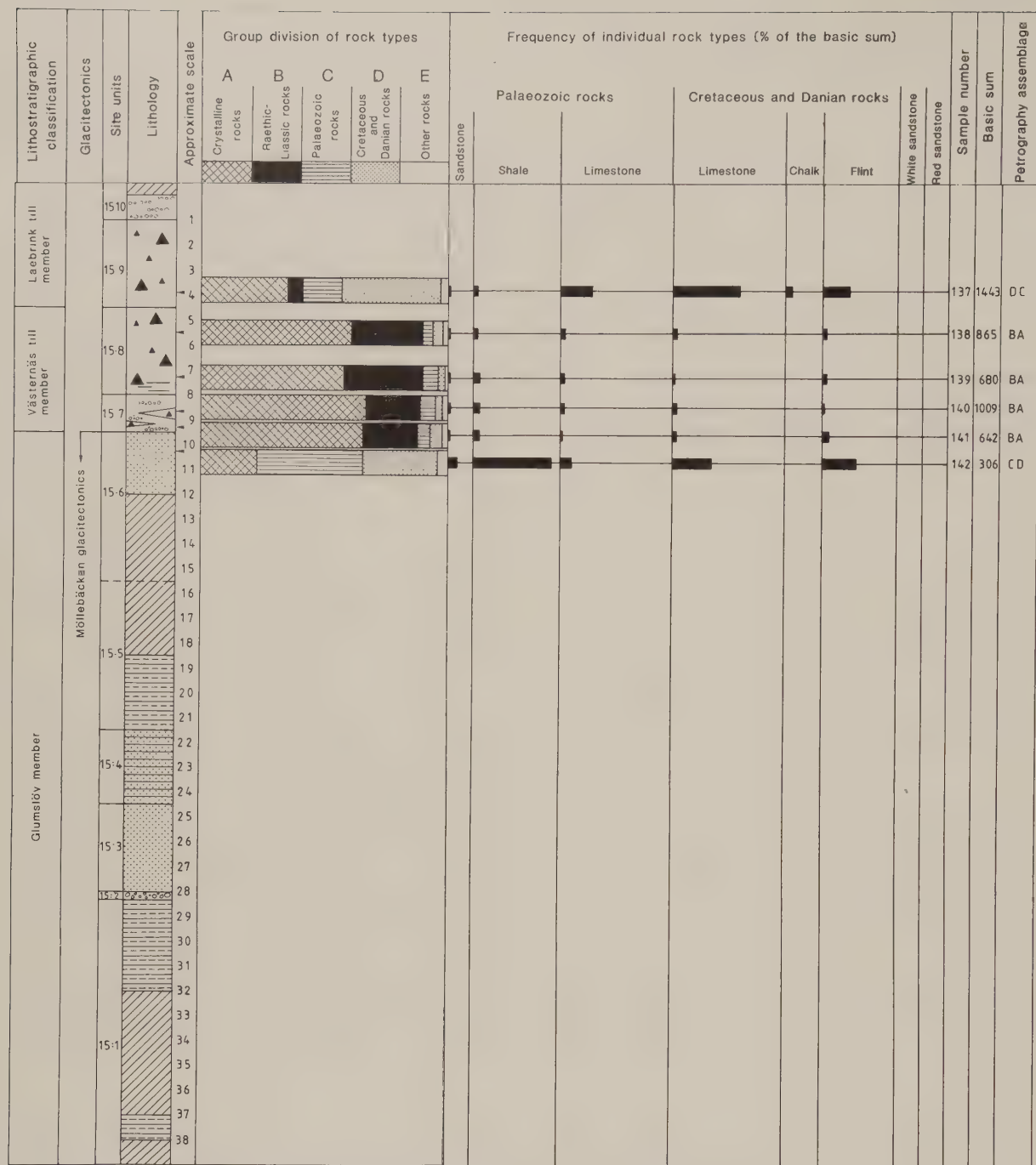


Fig. 55. Lithostratigraphy and petrography spectra, site 15.

the pit-wall was completely covered. The same diamicton was also found in another outcrop 100 m further to the west. In this outcrop the unit had a thickness of 3.5 m. A fabric analysis in the lower part of the unit shows a preferred orientation of clast longaxes at N10°W/S10°E.

Unit 15:10(Facies A and B). The diamicton in the western outcrop was covered by 1 m of stratified sand and fine gravel. The boundary between the two units was scoured.

Tectonics

Very few tectonic structures were seen in the outcrops, and the most striking evidence of dislocation was the general dip of the strata in the lower part of the section. Some of the bedding-planes were measured and were found to have a strike of N50–60°W and a dip of 20–32° towards the north-east. Two minor overthrusts were observed in the tilted sediments. One of them had a N47°W strike and a dip of thrustplane of 35° towards the south-west. The other had a N50°W strike and a dip of thrustplane of 40° towards the south-west. Both the bedding and the faults of the dislocated strata were truncated by units 15:7 and 15:8. The glacitectonics is shown in Fig. 56.

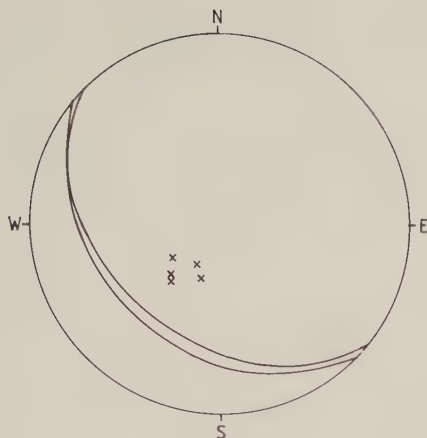


Fig. 56. Tectonics, site 15. Stereographic projection. Thrust faults are plotted as large circles. Bedding is plotted as poles to bedding surfaces on lower hemisphere.

5.1.16 Site 16

Site 16 (Fig. 57) extends from the mouth of the Hilleshög valley and about 900 m southwards, where the coast turns into the low coastal plane around the town of Landskrona. The cliff trends S30°E and is broken by some

deep ravines. The slopes are mostly covered by vegetation, but the strata can be studied in some outcrops at 16:200-305, 16:450-510 and 16:760-880. Dislocated strata, of the same type, are found in the outcrops. The sediments at 16:200-260 are strongly folded and thrust, and the most informative sequence of strata was found at 16:760-880. The litho-stratigraphic recording (Fig. 58) for site 16 is mainly based on the strata from this outcrop. Comparisons and correlations to the other outcrops in the cliff were made.

Lithostratigraphic units, site 16

Unit 16:1(Facies D). This unit constitutes the lower part of the southern outcrop but was also found in small cuttings just above the beach in other parts of the cliff. The exposed part of the unit has a thickness of 9 m in the southern outcrop, but the unit continues below the sea level. The silt and clay consists mainly of thick layers, 0.5 to 1.5 m, of homogeneous or convolute bedded silt, interlayered by clay laminae.

Unit 16:2(Facies B). A bed. of fine sand to coarse silt with convolute lamination and dish structures separates the sand facies from the lower silt and clay facies in the southern outcrop. The sand is approximately 10 m thick and contains tabular sets of cross-bedding, cross and planar lamination.

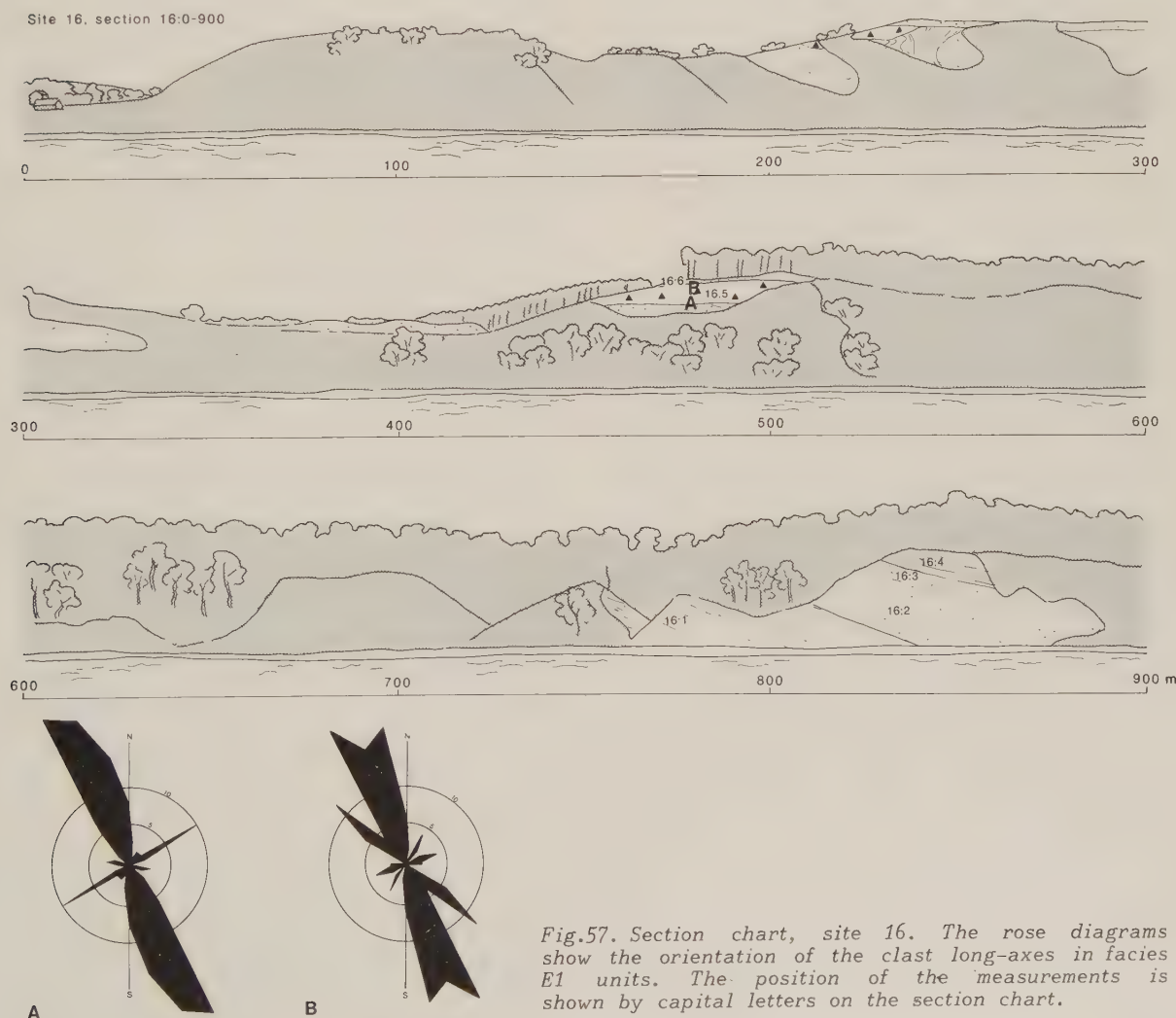


Fig. 57. Section chart, site 16. The rose diagrams show the orientation of the clast long-axes in facies E1 units. The position of the measurements is shown by capital letters on the section chart.

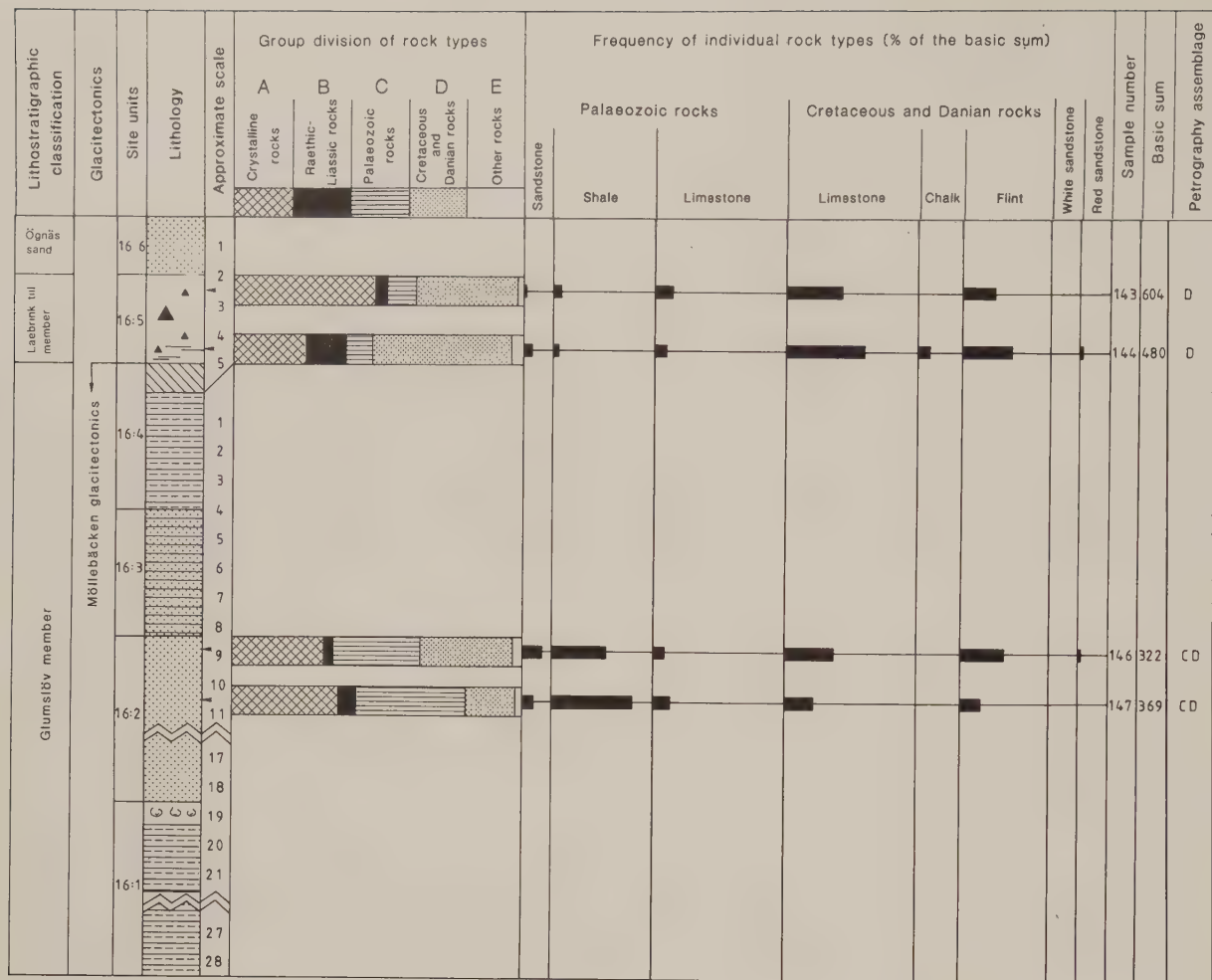


Fig. 58. Lithostratigraphy and petrography spectra, site 16

10 m thick and contains several tabular sets of cross-bedding, cross-lamination and planar lamination.

Unit 16:3(Facies C). The sand grades into alternating bed facies with repeated fining-upward sequences, separated by scoured surfaces. The unit is 4.3 m thick.

Unit 16:4(Facies D). The uppermost strata in the southern outcrop consist of more than 4 m of laminated silt and clay. The upper part of the unit is covered, and the total thickness is unknown. The unit was also found in the upper part of the outcrop at 16:200-305.

Unit 16:5(Facies E2f and E1f). The dislocated strata in the two northern outcrops are overlaid by a diamicton. The lower boundary of the diamicton truncates the bedding of the lower units. The basal part of the diamicton consists of a diffuse banding of silt, and the composition changes gradually upwards to a homogeneous diamicton with fissile structure. Two fabric analyses were made, one in the lower part and the other in the upper part of the 3 m thick diamicton at 16:475. The preferred orientation of clast longaxes was N25°W/S25°E in the lower part and N20°W/S20°E in the upper one.

Unit 16:6(Facies B). A sand unit can be traced on the top of the cliff in the northern part of the section. The sand ranges in thick-

ness from almost zero to a maximum of 2 m. The sand is a well sorted fine sand.

Tectonics

The deformation of the lower strata is most intense at 16:200-260, where the silt and clay laminae in the upper part of the cliff are vertically erected. Three fold-axes were measured in the complicated deformation. The fold-axes were in N62°E, N68°E and N70°E directions. The bedding-planes in the southern cliff-cutting have a general dip towards the south. Strikes and dips of bedding-planes were measured in the lower middle and upper parts. A clay lamina in the 16:1 unit had a N76°E strike and a 25° dip towards the south, a clay lamina in the 16:3 unit had a N68°E strike and a 21° dip towards the south, and a clay lamina in the 16:4 unit had a N84°E strike and a 30° dip towards the south. The measurements are shown in Fig. 59.

5.1.17 Site 17

The consulting firm VIAK AB was commissioned by the town of Landskrona in 1974 (VIAK AB 1974) in a groundwater investigation on the island of Ven. The boring was carried out 40 m to the north of the church in the central part of the island from a ground surface, situated 44 m a.s.l. Percussion

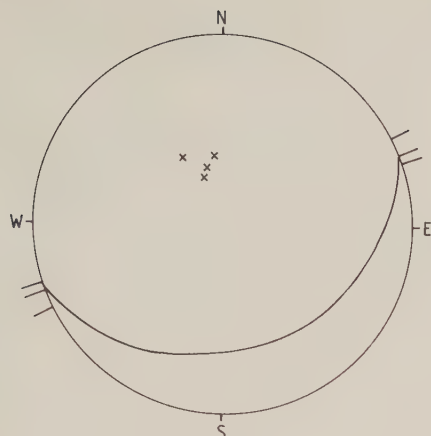


Fig. 59. Tectonics, site 16. Stereographic projection. Thrust faults are plotted as large circles. Bedding is plotted as poles to bedding surfaces on lower hemisphere. The trend of fold axes is shown by marks outside the circle.

drilling was made to a depth of 80 m, and the boring was continued by tube drilling down to 102.5 m (58.5 m a.s.l.), where the bedrock surface was reached. Disturbed samples were collected for each metre. The deposits were generally very consolidated, and the silt and clay sediments were partly brought up as fragments with the primary sedimentary structures still preserved. The lithostratigraphic sequence (Fig. 60) is based on the boring log which I recorded during the work. Repetitions of strata, due to overthrusts, cannot be excluded. However, a comparison between the strata in the boring and the general stratigraphy of the cliffs may make it possible to interpret such repetitions. Some of the following lithostratigraphic units may therefore belong to the same lithostratigraphic unit of the type succession for the area.

Lithostratigraphic units, site 17

Unit 17:1(Facies A), 100.5–102.5 m. The sediment on top of the bedrock surface consists of gravelly sand and sandy gravel.

Unit 17:2(Facies B), 79.3–100.5 m. The gravel is succeeded by 21 m of sand. A subdivision into three parts can be made. 99–100.5 m is dominated by fine sand, 92–99 m by medium sand and 79.3–92 m by fine sand. The sand is very well sorted, but contains larger fragments of amber, wood and other macrofossils.

Unit 17:3(Facies D), 75–79.3 m. At 75 m there is a rapid transition into silt and clay facies. Layers of coarse silt seem to dominate.

Unit 17:4(Facies E3D), 58–75 m. The unit consists of an alternation of silt and clay facies and diamicton facies. The main part of the samples had a diamictic composition with scattered gravel sized or larger clasts. The number of gravel clasts in each sample is too small for a statistical treatment of the gravel analyses.

Unit 17:5(Facies D), 54–58 m. The diamicton beds cease at 58 m, and the sequence continues as a silt and clay facies. Fragments of clay, probably representing distinct clay laminae, were brought up.

Unit 17:6(Facies B), 48–54 m. The silt and clay is overlaid by 6 m of fine to medium sand.

Unit 17:7(Facies C), 47–48 m. The sand samples from this level contain sediment clasts of red clay, which indicates a transition into alternating bed facies.

Unit 17:8(Facies D), 39–47 m. The material, brought up from this level, contains numerous sediment clasts of laminated silt and clay. The average content of silt is very high.

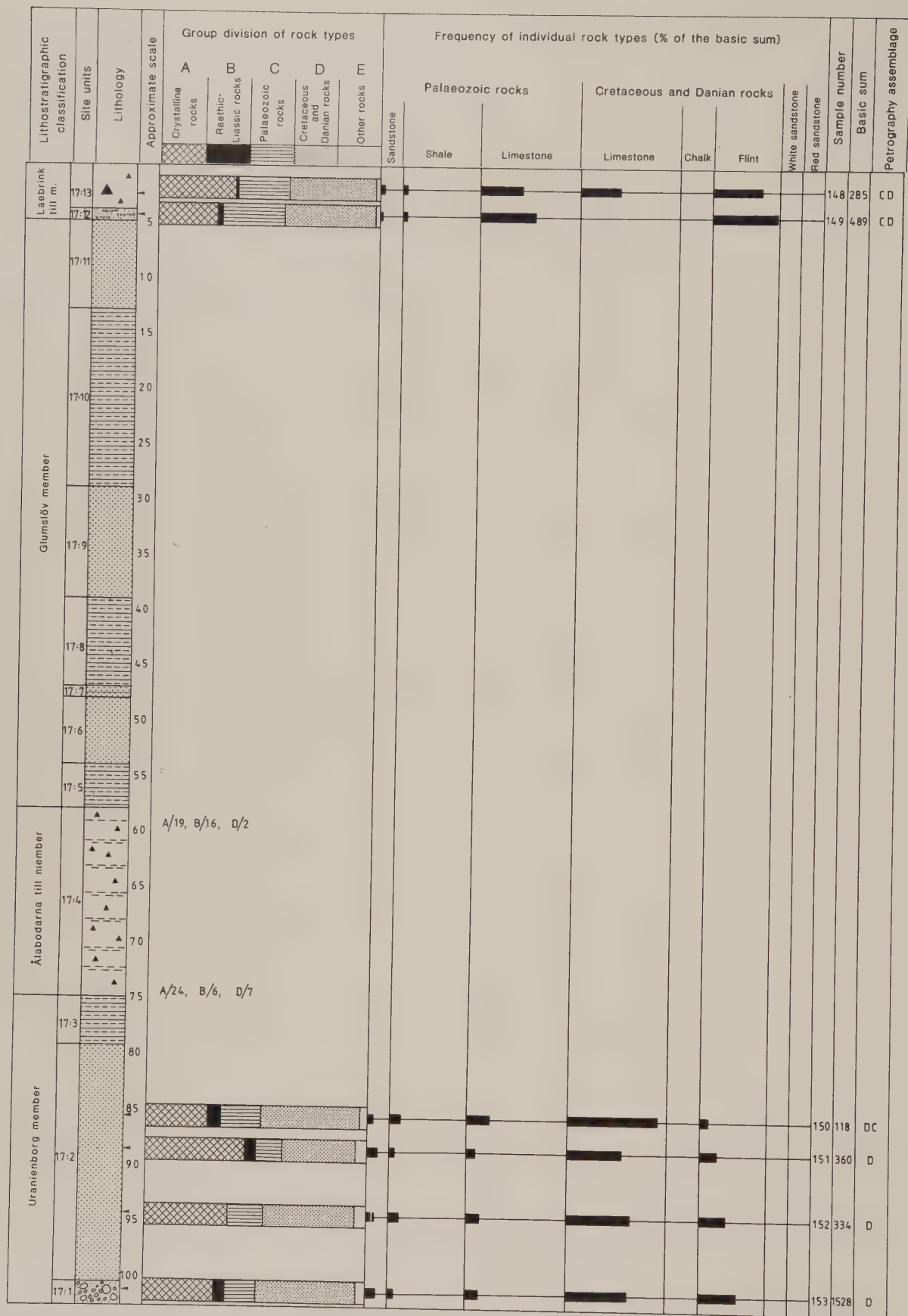
Unit 17:9(Facies B), 29–39 m. The sand unit is very fine grained with predominating very fine sand and coarse silt. Several sediment clasts with ripple lamination were brought up.

Unit 17:10(Facies D), 13–29 m. The samples from this level contain numerous sediment clasts with laminated silt and clay. The upper 5 m of the unit are more homogeneous, and the sediment clasts are structureless. Most samples contain between 70 and 80 % of silt.

Unit 17:11(Facies B), 5–13 m. The grain size composition increases upwards, and the samples from 5–13 m are dominated by sand and coarse silt. The clay content in bulk samples decreases from 15–25 % in the 17:10 unit to less than 5 % in the 17:11 unit.

Unit 17:12(Facies B), 4–5 m. At 5 m there is a rapid transition from the silty fine sand of unit 17:11 to a gravelly sand.

Unit 17:13(Facies E), 0–4 m. The upper part of the sequence is a diamicton, which also constitutes the top soil in the field around the bore-hole.



5.2 Lithostratigraphic correlations between the sites

5.2.1 Correlations based on stratigraphic position, facies states and unconformities

The first link in the lithostratigraphic correlation between the sites was a physical tracing of units between closely spaced exposures and by comparison of the stratigraphic position of units with correspondence in lithological character, Main Facies States. Lithological series, beds of different lithologies and thickness in a definite sequence, were also used in the correlation. The correlation of Facies E1 units was supported by the preferred orientation of clast longaxes and by the direction of striae on boulders in the diamicton units.

Angular unconformities were clearly recognized in most of the sections. Unconformities can be used as an objective basis for correlations (Thomas 1977, p 175). Two unconformities and two different structural directions were distinguished in the tectonics. Most of the major structures can be grouped in two main shear directions, one from the south-west to east sector and one from the north to north-east sector.

A grouping of strata into three groups was based on the unconformities and the directional patterns of the tectonics.

Group I: Subhorizontal units, not affected by any major glaciectonism.

Group II: Dislocated strata with shearing movements from E to NE.

Group III: Dislocated strata with shearing movements from N to NE and from SW to E or only from SW to E.

Angular unconformities were found between group I and group II strata, between group I and group III strata where group II was missing, and in occasional cases between group II and group III strata (Fig. 62). The correlation of units within group I and group III, based on the mentioned criteria, does not create any problems. The uncertainties are concentrated to the middle part of the sequence. Group II units are mainly exposed in the northern cliffs on Ven. The absence of deformation structures in the southern cliffs and on the mainland does not permit a correlation based on the unconformity between group I and group II strata. The undeformed units 20 and 21 may be correlated to the deformed unit 18 or may have a stratigraphic position just below unit 15. A correct position is impossible to determine and is not aimed at either, as the main part of the units 15 to 23 has a very restricted lateral extent and cannot be expressed as defined stratigraphic units for the whole area. The whole sequence, unit 15 to 23, can be regarded as a complex facies sequence, mainly of facies A, B and E, and the individual units having a restricted lateral extent. The upper and lower boundaries of the complex facies sequence are, however, generally well defined.

5.2.2 Comparison of directional properties in Facies E1 units

In this context the orientation of clasts is only used as a textural property which may characterize the Facies E1 units. The genetic and ice-flow directional implications will be discussed later.

The measures of clast longaxes orientation and striae directions on boulders in Facies E1 are compiled in Fig. 61. Only measures in homogeneous parts were made, and the few analyses in units 4 and 33 are due to the restricted distribution of the homogeneous facies in these units. The accordance between the few analyses within the two units is, however, good. Unit 15 and 22 cannot be distinguished by their directional properties. Both the units have boulders, striated from NE. Two trends are presented in the orientation of clast longaxes in unit 22. The orientations are at right angles to each other. These two directions are also recognized in unit 15.

Unit 28 has a distinct pattern of an upward change of orientation properties from southerly directions in the lower part to ESE directions in the upper part.

The significance of the directional properties can thus be summarized:

The directional properties support the correlations between sites of unit 28. The directional properties of units 15 and 22 do not distinguish the individual units but distinguish the two units from other units. The statistical significance of the directional properties of units 4 and 22 is based on far too few measurements and can thus not be confirmed.

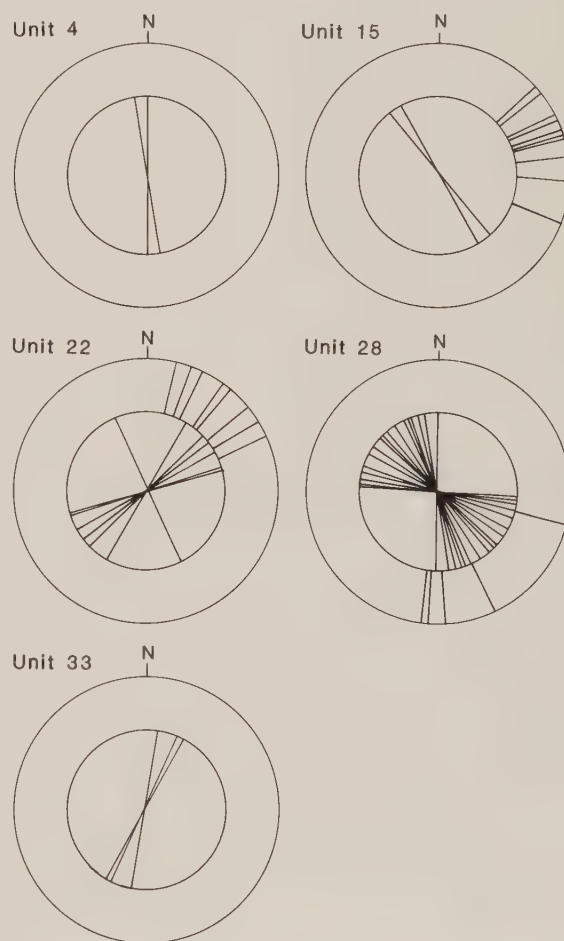


Fig. 61. Diagram showing directional properties in Facies E1 units. Striae are plotted in the inner circle and preferred orientation of fabric analyses in the outer circle.

Site	1		2		3		4		5	6	7	8		9		10		11		12		13	14	15	16	17	Lithostratigraphic classification		
	a	b			a	b	a	b				a	b	a	b	a	b	a	b	a	b								
GROUP I	B ¹⁴ _{0.4}	B ¹⁰ _{0.5}	B ⁶ _{0.4}	B ⁶ _{0.5}	B ¹² _{0.4}	B ¹³ _{0.5}				B ⁸ _{0.5}	B ⁸ _{0.5}	B ⁷ _{0.5}	B ⁷ _{0.5}	B ⁷ _{0.5}	B ⁹ _{1.2}	B ³ _{2.5}	B ⁸ _{1.1}		B ⁷ _{0.5}	B ⁸ _{0.5}					B ⁶ _{0.2}	34	Ögnäs sand		
	E ¹³ _{4.4}	E ⁹ _{5.7}	E ⁹ ₁	E ⁵ _{5.5}	E ⁷ ₅	E ¹¹ _{0.1}	E ¹² ₅			E ¹⁵ _{1.5}																	33	Kyrkbacken till	
																		A ⁷ _{0.1}									32	Laebrink till member	
	D ¹² _{0.4}	D ⁸ _{0.5}	D ⁸ _{4.5}	C ⁴ _{1.5}						D ⁷ ₂	D ⁵ _{0.7}	D ⁶ _{0.8}	D ⁶ _{0.8}	D ⁸ ₁				D ⁶ ₁			D ⁷ _{0.5}					31			
				E ³ ₃						E ³ _{0.5}	E ³ _{0.7}	E ³ ₁	E ³ _{0.5}	E ³ _{0.8}	E ³ _{0.2}				E ³ _{1.5}							30			
	A/B ¹¹ _{0.1}	E ⁷ _{0.3}	A/E ² _{0.1}	E ² _{0.6}	A ⁶ _{0.2}	A ¹⁰ _{2.2}	A ⁸ _{2.2}			A ⁶ _{0.1}													A/B ¹² _{1.3}	A/B ¹⁰ ₁		29			
	E ¹⁰ ₃	E ⁶ _{0.5}	E ⁶ ₇	E ¹ ₄	E ¹ ₃	E ⁹ _{0.3}				E ⁵ _{4.8}	E ⁵ _{0.2}	E ⁴ _{0.3}	E ⁴ _{2.3}	E ⁴ _{2.5}	E ⁴ _{1.2}	E ⁶ ₃	E ⁵ _{0.5}	E ⁶ _{1.8}	E ⁵ _{2.5}	E ⁶ _{0.5}	E ⁶ ₁	E ⁵ ₃	E ⁶ ₁	E ⁵ _{3.5}	E ¹³ ₄	28			
																			D ⁴ _{0.1}	C ⁵ _{0.9}	D ⁴ _{0.2}	D ⁵ _{0.3}					27		
	A/B ⁹ ₁																			A ³ _{0.2}	A ⁴ _{0.5}	A ⁴ _{0.2}	A ⁴ _{0.5}				26		
																					E ³ _{1.8}	E ³ _{2.5}						25	
																					B ² _{0.1}	B ² _{0.1}	B ¹⁰ _{0.4}				24		
						E ⁴ _{0.7}					A ⁴ _{0.2}																	23	Västernäs till member
			E ⁵ _{0.4}			E ¹ _{0.6}	E ⁷ _{0.5}	E ¹ _{0.5}		E ³ ₃		E ² _{0.2}		E ¹ _{0.1}									E ⁹ _{2.5}	E ⁸ ₃			22		
						A ³ _{2.5}	A ⁷ _{0.2}																	A/E ⁸ _{0.2}	E ⁷ _{0.14}			21	
						B ² ₂	B ⁶ ₆																					20	
																												19	
	GROUP II	E ⁵ _{1.5}									E ⁴ ₁	E ² ₂	E ³ _{1.5}															18	
										B ² _{4.3}	B ³ ₂	A/B ² _{0.1}	A ² _{4.3}	A/B ³ _{5.5}	B ¹ _{3.5}											17			
											C ² _{0.3}																16		
																											15		
											E ¹ ₁	E ¹ _{1.5}	E ¹ _{1.5}	E ¹ _{1.2}													14		
GROUP III																		B ² ₄	B ² _{2.6}		B ⁶ _{7.5}	B ⁷ ₇	B ⁸ _{4.6}				13	Glumslöv member	
																					C ⁵ ₅	C ⁶ ₂					12		
	D ⁸ _{10.5}		D ⁴ ₄				D ⁶ ₁₅	D ⁵ ₁₀										D ¹ _{12.5}	D ¹ ₆		D ¹ _{2.5}	D ⁴ _{2.5}	D ⁵ ₆	D ⁴ ₆	D ⁴ ₈	D ⁴ ₈	11		
	C ⁷ _{2.5}	C ⁴ ₃																			C ³ _{0.7}	C ³ ₂	C ³ ₃	C ³ _{4.3}	C ³ ₁		10		
	B ⁶ ₇	B ³ _{5.5}	B ³ _{2.5}			B ⁵ ₂	B ⁵ ₆	B ⁴ _{5.6}	B ¹ ₁												B ² _{4.3}	B ³ _{5.7}	B ³ ₄	B ² ₁₀	B ⁶ _{6.10}		9		
							A ⁴ _{0.2}																A ² _{0.02}				8		
	D ⁵ _{1.3}	D ² _{1.5}	D ² _{1.8}			D ⁴ ₃	D ² ₇	D ³ _{1.7}														D ² _{7.8}	D ¹ ₁₀	D ¹ ₉	D ¹ ₄		7		
	C ⁴ ₁						C ³ _{0.7}		C ² _{0.8}																		6	Ålabodarna till member	
	A ³ _{0.2}						B ² _{0.6}																				5		
	E ³ _{1.8}	E ³ ₆			E ³ _{1.1}	E ³ _{1.3}	E ³ _{1.5}	E ³ ₇														E ³ _{2.1}	E ³ ₄				4		
	E ² ₂		E ¹ ₃					E ² _{1.5}											E ¹ ₁		E ¹ ₂	E ¹ ₁					3		Uranienborg member
																										D ³ _{4.3}		2	
																									B ¹ _{2.1}		1		

5.2.3 Correlations based on the gravel analysis results

The petrographical composition of a sediment can be interpreted from several aspects:

(a) It may be used as a possibility to identify a stratigraphic unit, (b) the clastic particles may provide evidence of provenance which in turn can be used to infer ice flow directions and (c) the petrographical composition may also reflect the mode of debris entrainment, transport and deposition.

The possibility to identify a glacial sediment by its petrographical content is dependent on the homogeneity or regulated variety in the sediment. The constancy of different parameters in tills was investigated by Bahnson (1973) in some till sequences in Denmark. He found a constancy in petrography, grain size distribution and chemical content within the individual tills and considered it possible to differentiate tills in a boring or section by these properties. May and Dreimanis (1976) stated, by a statistical treatment of the vertical variability of different parameters in two till sequences of Catfish Creek Drift, that the petrography of clasts tended to be relatively more variable than carbonate and granulometric parameters.

Regulated vertical changes were found in Bräcke till in north-west Skåne (Lagerlund 1971, 1977b), and they were interpreted as a turning of the ice flow direction during the deposition of the till. Vertical variations in a glacier were described from Svalbard by Boulton (1970). The vertical changes were the result of a successive basal incorporation of different rock types by a net basal freez- ing. A preservation of the vertical variations can be expected in the till if deposited by basal melt-out processes from stagnat ice. A possibility to verify and supplement the preliminary stratigraphic correlations between the sites and to find out the petrographical homogeneity of the individual units was achieved by a statistical treatment of the gravel analysis results.

A classification of the samples, based on their petrographical composition, was made by a simple statistical treatment of the total variance of the samples. The petrography spectrum of each sample was classified according to its deviation from the mean of the four rock groups A, B, C and D, counted on all the samples. The capital letter of the rock group or rock groups with a positive deviation in the sample was used in the classification. The rock group with the most pronounced deviation was put first. Ten combinations of the four rock groups were represented in the 153 samples: A, AB, BA, B, C, CD, DC, D, CA and DCA petrography assemblages.

The distribution of the petrography assemblages in the individual stratigraphic units is shown in Fig. 63.

A zonation of the lithostratigraphic record can be made from the vertical distribution of the petrography assemblages. The zonation is based on the grouping of the petrography assemblages into one group with combinations of the A and B rock groups and another group with combinations of the C and D rock groups. The boundaries between the different

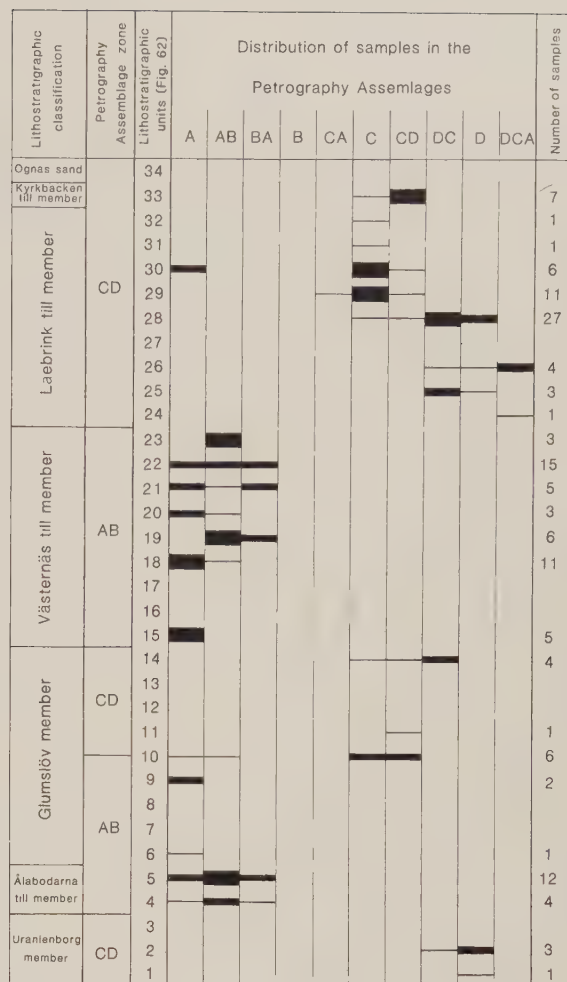


Fig. 63. The distribution of petrography assemblages in the lithostratigraphic units. A thick line means $\geq 50\%$ of the samples, or in units with less than 6 samples, ≥ 3 samples. The intermediate line means $30-50\%$ of the samples or 2 samples. The thin line means $< 30\%$ or 1 sample.

petrography assemblage zones represent distinct transitions from one type of petrography composition to another.

The petrographical homogeneity of the units can also be visualized in a computer diagram where the mutual relations between the 11 petrographical variables are considered. The diagram was achieved by plotting the samples on the first and second principal components based on the covariance matrix. The plot of samples are related to the different stratigraphic units of Fig. 64, and the petrographical accordance between samples from the same stratigraphic unit can be interpreted from the clustering of plots with the same figure, symbol. The position of the petrography assemblages is shown in the diagram.

The samples from the same lithostratigraphic

Fig. 62. Correlation chart showing the lithostratigraphic correlations between the sites. The squares show facies state in letters (see chapter 4.2 for description), the number of unit in the site descriptions (upper right), and the average thickness (in metres) of the unit in the section (the lower figure in the square).

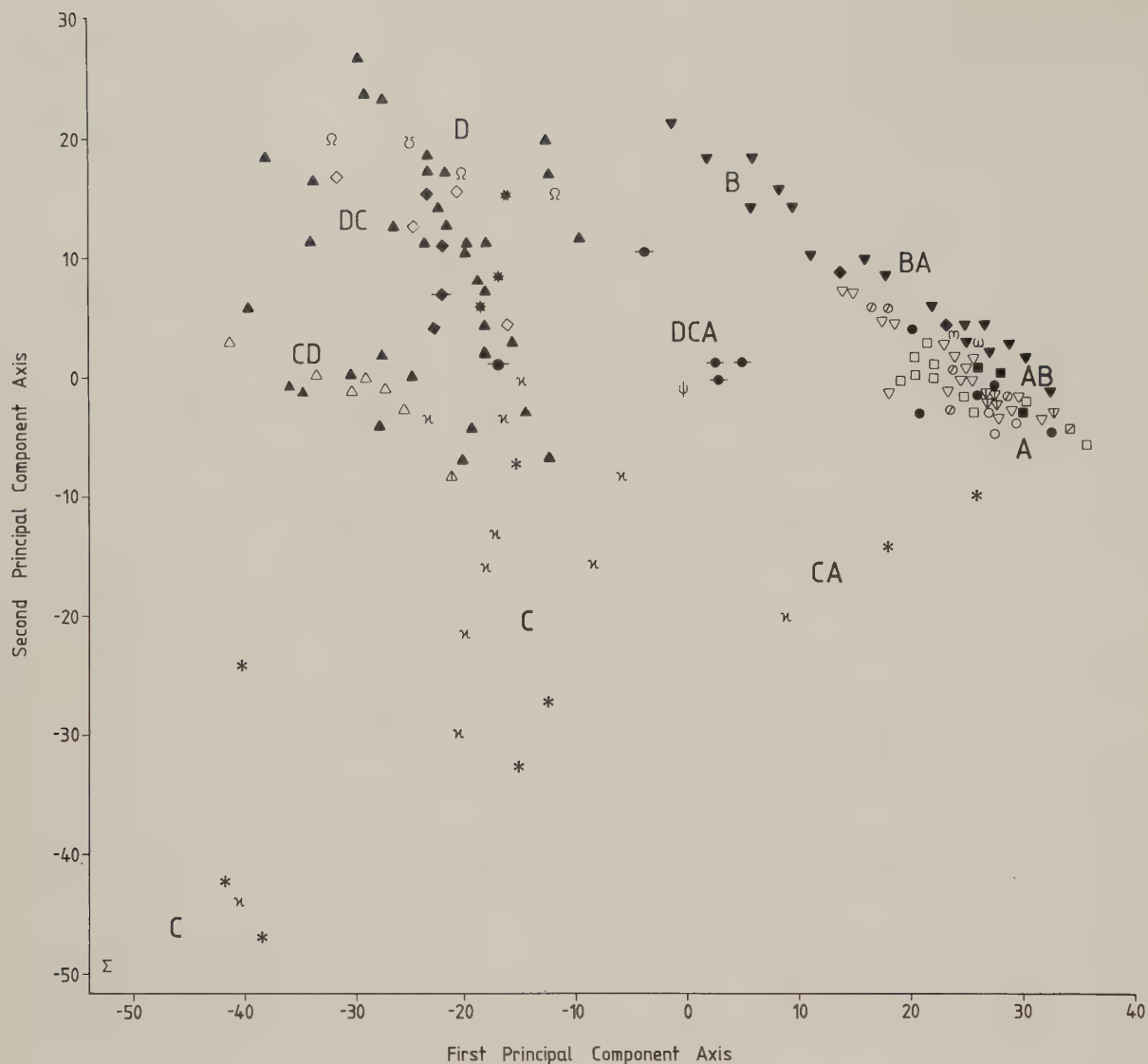


Fig. 64. Comparison of the petrographical composition of samples from the different lithostratigraphic units in Ven-Grumslöv area. The figure of the units (from Fig. 61) and corresponding symbols are shown below the diagram. The samples are plotted on the first and second principal components and related to the petrography assemblages. All the samples included in the numerical analysis are shown in the diagram. A few of them, belonging to units 28 and 30, were not accounted for in chapter 5.1. Some of these samples will be discussed in chapter 6.6.2.

unit are generally very well clustered. This indicates a great homogeneity in the petrographical composition of clasts in the main part of the units. Two main units, unit 10 and 30, show a scattered distribution of samples in the diagram. The differences in petrographical composition of unit 10 is related to the vertical position of the sampling points.

The AB and BA petrography assemblages occur in the lower part of the unit, and the C and CD petrography assemblages occur in the middle and upper part of the unit. The differences in unit 30 are related to variations between individual thin beds in the whole unit. The variations within the two units are a repeated and characteristic pat-

tern, which can be used in the identification of the units. The provenance and genetic implications of the variations will be discussed later.

5.2.4 The reconnaissance of reworked rock clasts

The petrography of the sedimentary particles in the glacial sediments are dependent on the bedrock and sediments, overridden by the glacier. The bedrock mosaic in Skåne and surrounding areas is shown in Fig. 65. The bedrock is mainly covered by Quaternary deposits. The presence of Quaternary deposits, containing multiple till units with



Fig. 65. The bedrock distribution in Skåne and eastern Sjaelland, simplified from Bergström and Shaikh (1980, 1982), Bergström et al. (1982) and Sorgenfrei (1971).

different petrographical composition, has been documented from numerous borings and excavations in Skåne. The supply of older sediments available for erosion, must have been comprehensive, and mixing of material during the divergent ice movements in Skåne and Denmark may have resulted in petrography contents, which do not necessarily reflect the bedrock, overridden by the glacier. This source of contamination was discussed by Marcussen (1973) in a critical review on the use of indicator boulders. The contamination of rock types from older deposits can, according to Lagerlund (1971, 1977b), be accounted for by a method of interpretation where the spectrum of the petrography composition from a certain till in an investigated section is compared to the immediate lower unit in the same section, and only the new additional rock types are regarded as indicators of the ice flow direction. That means that the knowledge of the

stratigraphic position and the petrographical composition of the older deposits in the area is necessary for the interpretation. The distribution of petrography assemblages in the individual stratigraphic units (Fig. 63) and the clustering of samples in Fig. 64 show a close association of rock types from rock groups A and B (variables 1, 2 and 9 in the computer program) and from rock groups C and D (variables 3, 4, 5, 6, 7, 8 and 10 in the computer program). The two main clusters in the diagram (Fig. 64) are distinctly separated. The cooperation of rock types can be shown by the loadings of variables on component 1, which gives the mutual relations between the petrographic variables in the matrix. The first principal component accounts for 66 % of the total variance. The loadings show a cooperation of variables 1, 2, and 9 and a cooperation of variables 3, 4, 5, 6, 7, 8 and 10 (Fig. 66). This can be compared to the bedrock distribution

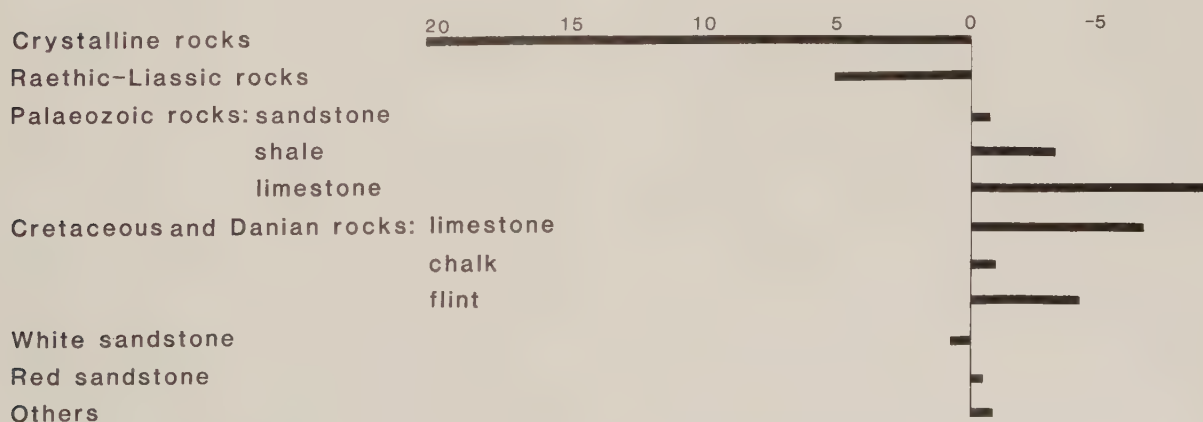
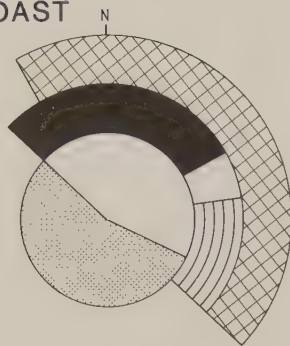


Fig. 66. The cooperation of the different petrographical variables shown by the loadings of variables on component 1. Rock types with negative values are cooperating and rock types with added values are cooperating.

THE MAINLAND COAST



VEN

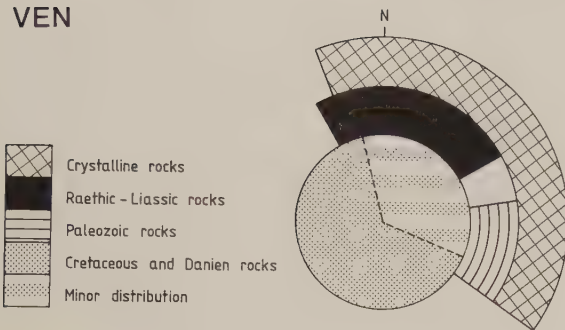


Fig. 67. The inferred transport direction of the main rock groups in samples from the mainland coast and from Ven.

of the different rock types (Fig. 65). Fig. 67 shows the most likely transport direction inferred from the bedrock in situ. The direct accordance between the mutual relation of different rock types in the same samples and the inferred transport direction is obvious and can be interpreted as an evidence of a provenance of clasts directly from a bedrock source.

The contamination of reworked rock components can also be interpreted from the mutual relations between the petrographic variables. A petrography spectrum of a sample, containing non-cooperating rock types, gives an indication of rework from older deposits. The mixing of rock types from different directions in the samples is generally very restricted, and the petrography spectra are mostly clean.

Minor amounts of non-cooperating rock types are, however, present in some of the samples. The samples from units 16 to 25 are dominated by rock groups A and B. Most samples also contain a minor amount of non-cooperating

rock types from rock groups C and D. The total share of the two rock groups varies between 3 and 18 % in the samples. The ratio between the two rock groups are, however, similar in all samples. Rock group D is dominating in 40 out of the 46 samples. This repeated pattern of composition indicates that the older reworked sediment was dominated by rock type D and that the content of rock type C was slightly less.

5.2.5 Comparison of grain size composition of Facies E1 units

The grain size composition is another criterion which can be used to differentiate, describe and identify tills (May and Dreimanis 1976). Krumbein (1933) recognized that the grain size of individual tills is more or less homogeneous. Textural variations were also successfully used as one of the parameters in the identification and correlation of eight different till units in the Danville area, Illinois (Johnson et al. 1971).

A comparison of the grain size composition of the homogeneous diamicton facies (Facies E1) in the Ven-Glumslöv area was made. Between 18 and 40 samples were taken from units 4 and 5 (about half of the samples were taken in 0.5 to 1 m thick homogeneous beds in unit 5) from unit 16, unit 23 and unit 29 (unit numbers from Fig. 61). The average percentage distribution of sand, silt and clay and the standard deviations (SD) are displayed in Table 6.

The variation of sand, silt and clay ratio within the individual units is generally moderate, and the magnitude of standard deviation is not larger than the grain size distribution can be used to describe and characterize the different units. The variations between some of the units are, however, too small to differentiate the units. The composition of units 15, 22 and 28 are too similar to allow an indication, based on the grain size composition.

A combination of grain size composition and petrography of clasts in the diamicton gives a more unambiguous possibility for the identification of the different homogeneous diamicton units. The diamictons of the AB assemblage group are easily distinguished from diamictons of the DC assemblage group even by very simple field investigations. The two diamicton units 15 and 22 are almost identical in grain size and petrographical composition. They are, however, easily distinguished from units 4 and 5, which also belong to the AB assemblage group, by a distinct difference in grain size composition.

Table 4. The percentage mean and standard deviation for sand, silt and clay contents in the homogeneous diamicton units (units 4-5, 15, 22 and 28) in the Ven-Glumslöv area.

Lithostratigraphic unit	sand (-1 to 4 Ø) M (SD) %	silt (4 Ø to 9 Ø) M (SD) %	clay (>9 Ø) M (SD) %	n =
unit 28	47.08 (3.90)	36.13 (3.38)	16.79 (3.02)	40
unit 22	51.25 (7.01)	31.69 (5.00)	17.06 (3.25)	24
unit 15	50.07 (5.92)	33.69 (4.92)	16.24 (3.85)	18
unit 4 - 5	21.7 (5.25)	46.8 (2.94)	31.5 (6.48)	38
units 4-5 (from the mainland)	17.71 (4.01)	46.0 (2.00)	36.29 (4.87)	18
units 4-5 (from Ven)	24.83 (3.97)	47.39 (3.45)	27.78 (5.74)	20

5.2.6 The correlation of the boring log to the lithostratigraphy of sites 1 to 16

The correlation of the log from the VIAK boring on Ven to the strata in the investigated sections was based on stratigraphy, facies states, thickness of strata, lithological series and petrographical composition.

The uppermost unit, unit 17:13, was correlated to unit 28 in Fig. 61. The correlation was based on stratigraphic position, grain size composition and petrographical composition. The thick strata of Facies B, C and D (units 17:5-17:11) have their only equivalents in units 6 to 14. The repetition of strata due to glaciectonic thrusting, cannot be evaluated from the boring log and the exact correlation of the units can consequently not be made. Unit 17:4 has a very characteristic grain size composition with a high silt and clay content and dispersed clasts. The cumulative curves from the grain size analyses are almost identical to curves from samples of the Ålabodarna member, taken in the cliffs on Ven. The clast content in the samples from the boring is generally very low, and a statistical treatment of the identification of rock types was not possible. The clasts are of the A and B rock groups.

5.3 Lithostratigraphic classification

The lack of tradition in lithostratigraphic classification of Quaternary deposits in Europe has been emphasized by Bowen (1978), and the chronostratigraphy has often influenced the definition of lithostratigraphic units (Luttig et al. 1969). A strict lithostratigraphic classification has since long been used in U.S.A. (e.g. Willman and Frye, 1970), and during the last years, after publication of the International Stratigraphic Guide (Hedberg 1976), an increasing number of European Quaternary stratigraphers have started to use the international recommendations for stratigraphic classification. A formal lithostratigraphic classification of Quaternary deposits in Skåne was introduced by Lagerlund (1980).

The general consideration in lithostratigraphic classification is lithological similarities within the units defined (Hedberg 1976). The strata in glaciated areas often show very large variations in lithological character, and the individual beds often have a very restricted lateral extent. Lateral thinning and gradual passing into a different facies state are frequent. Intraformational glaciectonic deformation often makes the lateral tracing and mapping of units difficult. The basic unifying feature in glacial sequences is often a complex pattern of facies relationship between different types of sorted and diamictic sediments. A laterally extensive, more or less homogeneous, diamicton (i.e. a basal till) is frequently recognized in a facies complex, and can then be used as a marker bed of the unit defined.

The designation till is generally used as a prefix to the unit. This may stand for the diamictic composition of the sediments but also for a general lithological complexity of the glacial sequences.

The choice of hierarchy is governed by the lateral and vertical extent of the unit defined. A formation should be capable of being

mapped as a tabular body of strata (Hedberg 1976). The extensive glaciectonic deformation and the lateral discontinuity of some of the units in the area makes the recognition and mapping difficult in some parts of the area. The different till units are therefore given a member status. This choice of hierarchy follows Willman and Frye (1970) and has also been used for corresponding units in other parts of Skåne (Lagerlund 1980, Berglund and Lagerlund 1981). The grouping of strata into lithostratigraphic units was based on similarities in lithological character or similarities in the pattern of facies relationship, similarities in the pattern of petrographical composition and evidence of unifying directional patterns (kinostrostratigraphic grouping according to Berthelsen 1973, 1978).

Uranienborg member

The lower units, units 1, 2 and 3 (Fig. 62), were only studied in the boring on Ven. They consist of sorted sediments, and the member is dominated by the 21 m thick stratum of well sorted fine to medium sand.

The petrography spectra of fine gravel samples belong to the D and DC petrography assemblages. Clasts of amber and twigs and fragments from Tertiary conifers are characteristic for the sand. Palynomorphs are rare in the sand (pers. comm. Urve Miller).

The Uranienborg member rests on bedrock of Cretaceous age. The unit is named from the historical monument of Tycho Brahe, which is situated close to the boring site.

Ålabodarna member

The member comprises units 4 and 5 (Fig. 62) and is approximately 20 m thick. The sequence is predominated by stratified or crudely stratified diamicton with silt and clay intrabeds. The silt and clay content in the diamicton beds is very high and generally exceeds 70% by weight. The petrography spectra of samples from the member belong to the AB, BA or B petrography assemblages.

The lower boundary of the unit is defined as the distinct change into a diamictic silty clay at -31 m in the boring on Ven. The change in texture is accompanied by a change in petrography composition.

The unit is named from the fishing village of Ålabodarna on the mainland coast.

Glumslöv member

The member consists of the sorted sediments of units 6 to 14.

The strata are fine grained, facies B, C and D, except for some thin units of gravel, resting on scoured surfaces. The unit is characterized by two fining upward sequences with a total thickness of approximately 20-25 m on Ven and a 7-10 m thick basal unit of facies D, interrupted by a scoured surface or a rapid change into facies B, followed by a 7-15 m thick fining upward sequence, which in turn is followed by a coarsening upward sequence on the mainland.

The petrography composition of the lower part of the unit belongs to the AB and BA petrography assemblages and to the C, CD and DC petrography assemblages in the upper part. The change in petrography composition is not connected to any distinct lithological boundary and therefore the lower boundary of the unit is defined as the distinct change in texture between the diamictic sediments of the Ålabodarna member and the stratified and laminated sorted sediments of the Glumslöv member.

The upper boundary of the unit is a disconformity.

mity due to a glacitectonic deformation of the strata and an erosional hiatus. The unit is named from the village of Glumslöv on the mainland.

Västernäs till member

The member is characterized by diamicton, gravel, sand and a minor content of alternating bed facies and silt and clay facies. The individual beds generally have a very restricted lateral distribution, and the units cannot be traced in the whole area. The thickness of the member ranges from zero to approximately 7 m. The member comprises both deformed strata from group II (Fig. 61) and subhorizontal strata from group I. The units 15 to 23 are included in the Västernäs till member.

The lower boundary is an angular unconformity to the lower units. The petrography composition is characterized by the A, AB and BA assemblage groups.

The member is named from the north-west cape, Västernäs, on Ven.

Laebrink till member

The member consists of units 24 to 32 from the Group I strata.

Unit 28, Facies E1, has an extraordinary lateral continuity and is considered a marker bed for the member. The strata below the marker bed have a restricted lateral distribution and are dominated by facies E3B and facies A. The strata on top of the marker bed consists of facies E3B or facies A at levels more than 13 m a.s.l. and facies E3D at levels below 10 m a.s.l. The uppermost unit of the member is of facies D or facies C.

The petrography composition of the samples from the member is of the C, CD, DC, D, CA or DCA petrography assemblages.

High ratio of rock group C is characteristic of samples from the upper part of the member, and units 29, 30 and 31 may display extreme percentages of palaeozoic limestone. Some of the samples from unit 30 belong to the A assemblage group. Those samples contain a very specific type of fine grained red granite, which often occurs together with red porphyry with black quartz grains from the Baltic.

The lower boundary of the member is mostly

represented by the distinct basal surface of the marker bed (unit 28). The boundary is generally represented by a distinct change of textural and petrographical properties. The member has got its name from the low cliff of Laebrink on the north coast of Ven.

Kyrkbacken till

Kyrkbacken till consists of one single unit, which can be traced as a 3 to 5 m thick heterogeneous diamicton in the upper part of the cliffs from Kyrkbacken village on the west-coast of Ven to the cliffs in the southern part of the island. The unit is characterized by textural inhomogeneity and lateral changes in sedimentary structures. The different structural types of facies E (facies Elf, facies E2 and facies E3) are found within the same unit.

The petrography composition of analysed samples from the unit belongs to the CD and C petrography assemblages. The samples do not fully represent the complexity of the unit as smudges of chalk, and chalky diamicton and slabs of reddish diamicton with high ratio of red sandstone (rock type 10) were avoided during the sampling. The main part of the samples have, however, a characteristic high ratio of rock type 10.

The unit is named from the village of Kyrkbacken on the west-coast of Ven.

Ögnäs sand

The uppermost stratigraphic unit is the Ögnäs sand. The sand bed is generally about 0.5 m thick but increases in thickness in the cliffs to the north of the Kyrkbacken village. The unit is discontinuous but is present in almost all the sections. The distribution in the cliffs gives a misleading idea of the general vertical distribution, as the sand is usually restricted to a narrow zone of uncultivated area nearest to the steep. The sand is characterized by a diffuse humic banding and a grain size composition of fine to medium sand. A lower part of gravelly sand was distinguished in some parts of the northern cliffs, but the destruction by soil processes made a separation into individual beds very difficult.

The sand is named from Ögnäs, the south-west cape of Ven.

6 Stratigraphic synthesis, sedimentary processes and palaeoenvironmental reconstruction

The second step in the investigation comprises the interpretation of sedimentary processes and environmental conditions. A prerequisite for a reconstruction of a palaeoenvironmental history is a detailed lithostratigraphic study, which makes it possible to put the genetic interpretations in a well defined stratigraphic relation.

The correlation chart in Fig. 62 was used as the basis for the studies of the changes of the physical environmental conditions. The chart shows a succession of 34 defined units, grouped in five members and two beds. The lateral distribution of the individual units can also be interpreted from the chart.

The descriptions and interpretations of the sediments follow the succession of stratigraphic units in the correlation chart and start with the lowest units, exposed in the cliffs, the Ålabodarna member.

The Uranienborg member was only registered from the boring, and the boring data give restricted information, which can be used for a palaeoenvironmental reconstruction. The Uranienborg member has its main importance as a unit which can be used in the regional correlations and will be discussed in chapter 7.

6.1 Ålabodarna member: gravity flow sedimentation in a proglacial lake.

The approximately 20 m thick Ålabodarna member is completely dominated by Facies F3D. A rough estimation of the percentage distribution of different sediment types in the stratified facies is: diamicton beds 75 %, laminated silt and clay 15 %, and silt with clay intraclasts 10 %. A fissile facies, Facies E1f or E2f, was found in some of the sections and has its stratigraphic position in the lowermost part of the member.

The whole member is very fine grained. The mean size of the diamicton samples ranges from 5.1 to 8.1, and the gravel ratio is extremely low and does not exceed 5 % in any of the analysed samples. Large clasts are very rare, but a few scattered boulders are generally present in the exposures. Some of them are up to 1 m³ large. Fig. 68 shows the grain size composition of samples from homogeneous diamicton beds from different parts of the area. The samples from the boring are not connected to a definite facies, as structural studies in the disturbed samples

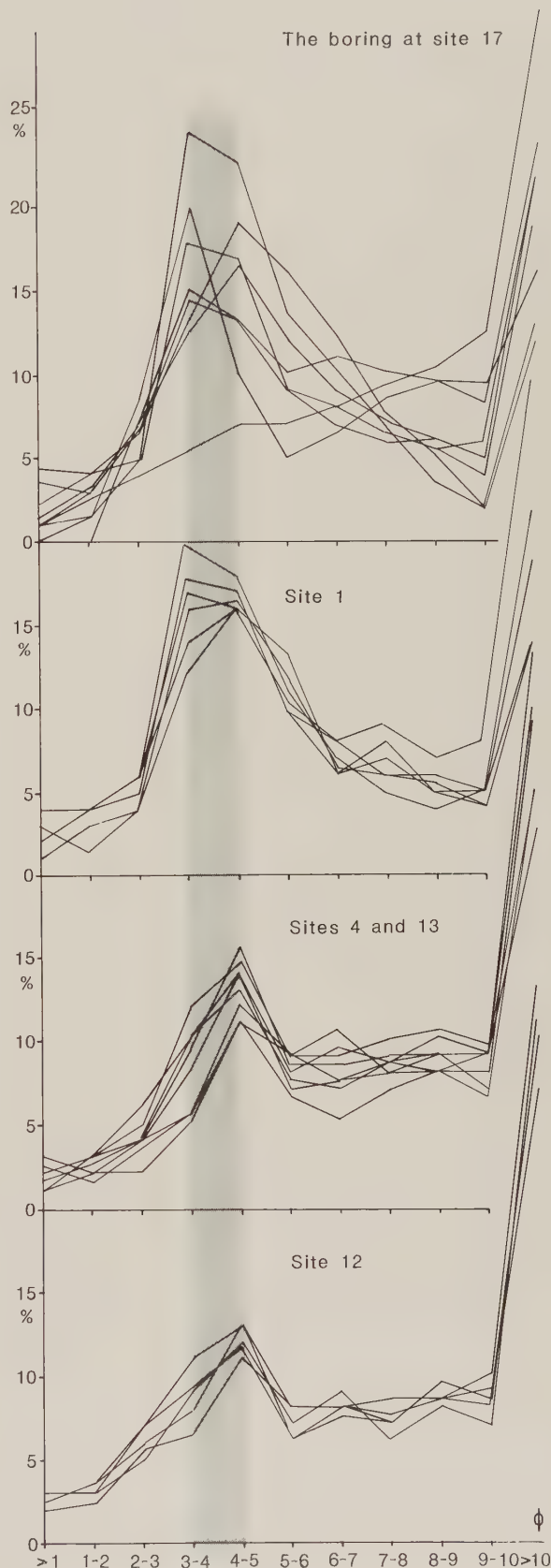


Fig. 68. Grain size distribution of samples from homogeneous diamicton beds of the Ålabodarna member from different parts of the investigated area. The diagram from the boring at site 17 also includes samples of other facies as a distinction between E1, E2, E3 and D facies cannot be made from the disturbed samples.

were not possible. Some of the samples can be compared to samples from the west-coast of Ven.

The samples show striking similarities both within and between the different parts. On the mainland and on the south coast of Ven the diamictons have their primary mood in the 4-5 Ø size grades. In samples from the boring and from the west coast of the island the primary mood is contributed by the 3-4..Ø size grade.

The laminated sediments generally have a very high silt and/or clay content. Only some samples from cross-laminated and parallel-laminated silty parts show mean sizes larger than 5.8 Ø. None of the samples has a mean size larger than 4 Ø.

6.1.1 Provenance of material

Dispersed fragments of molluscs are found in the diamicton layers. Marine foraminifera are also present (Rasmussen 1973). The mollusc fragments show a variety of species, representing different climatic conditions (table 5). The shells are broken and abraded, some of them even striated.

The mixing of species with different demands for water temperature, and the fact that the shells are broken and abraded indicate a rework and redeposition of older marine sediments.

All the species enumerated in table 5 are frequent in Eem and Weichselian marine sediments in the Kattegatt area, as shown from the Skaerumhede borings (Jessen et al. 1910, Bahnson et al. 1974). Several of the species are distinctly separated in the Skaerumhede stratigraphy, as *Turitella communis*, belonging to the lower sediments of Eemian interglacial age, and *Portlandia arctica* in the upper part, representing the arctic conditions during the Weichselian glacial age.

Table 5. Mollusc species from the Ålabodarna member. The classification of the mollusc fragments is made by Kaj Strand Petersen.

<i>Portlandia arctica</i>	<i>Mya truncata</i>
<i>Portlandia lenticulata</i>	<i>Hiatella arctica</i>
<i>Gardium ciliatum</i>	<i>Arctica islandica</i>
<i>Astarte borealis</i>	<i>Turitella communis</i>
<i>Leda pernula</i>	<i>Buccium sp</i>
<i>Leda minuta?</i>	<i>Dentale entale</i>
<i>Macoma calcarea</i>	

The combination of rock types in the gravel samples from the diamictons (see chapter 5.2.3) also supports a provenance of the material from areas to the north of the studied area. The high silt and clay content in the sediments can thus be related to the source material and does not necessarily reflect specific depositional and environmental conditions.

6.1.2 Sedimentary characteristics and interpretations

The fissile diamicton facies

Sedimentary characteristics

Two types of fissile diamicton were found in the lower part of the member, a massive fissile diamicton and a crudely laminated

fissile diamicton.

The massive diamicton is exposed in a 17 m long section at site 12. The lower strata in the cliff are affected by glacial tectonism, and the lower boundary of the diamicton is a thrust plane. The foot wall consists of sand and silt from the Glumsløv member. The upper boundary is an erosive contact. The exposed part of the diamicton has a maximum thickness of 3.5 m.

The diamicton can be separated into indistinct tabular units, 30 to 80 cm thick. The boundaries between the units are visible only when the sediments are dried up and are developed as widely spaced parallel and subhorizontal fissure planes with slickensides. A fissility of second order can be seen in the space between the parallel boundaries. This fissility pattern consists of low angle fissures which split the diamicton into irregular lenses (Fig. 69). The fissures are not equally distinct. Some of them are more consistent, and slickensides have been developed on the surfaces. Others are more indistinct and lack slickenside structures.

The grain size composition of 5 samples is displayed in Fig. 68. The sampling spots were scattered over the section. The composition of the samples is very similar, and the percentage distribution of the different Ø size grades varies within a few percentages. The diamicton is regarded as homogeneous in texture.

The crudely stratified fissile facies is exposed in the eastern part of section 4 (4:375-400). The upper part of the approximately 2 m high exposure consists of a crudely

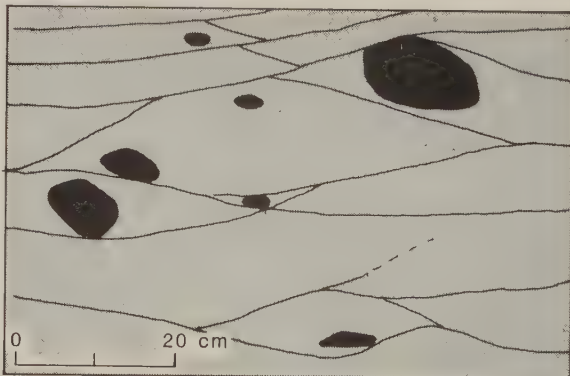


Fig. 69. The second order low-angle fissility pattern in facies E1f at site 12.

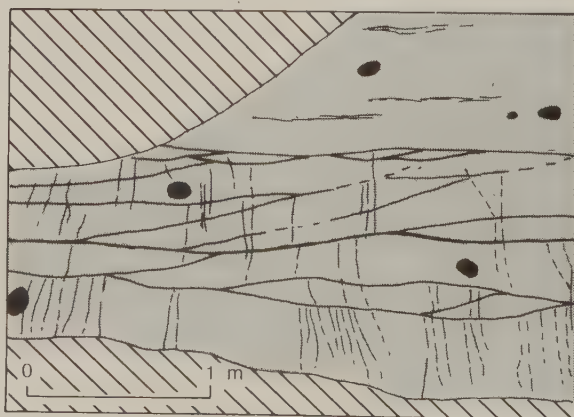


Fig. 70. Low-angle fissility pattern and vertical joints in facies E2f at site 4.

stratified diamicton, and the lower part is of the crudely stratified fissile facies. There is a distinct boundary between the two different facies.

The fissile diamicton is transected by a combination of vertical and low-angle fissures (Fig. 70). The low-angle fissures conjugate and form flat-lying irregular lenses. The fissures are generally 20 to 30 cm apart. Slickenside structures were recognized only on some of the surfaces and were especially well developed on the almost horizontal wavy surface in the middle of the picture (Fig. 70). The vertical fissures are widely spaced in the upper part but become a close pattern of columnar joints in the lower part (Fig. 71).

The fissile diamicton is dark grey and contains diffuse zones of thin light grey streaked-out slabs and lenses of very fine sand and silt. It also contains some zones with dispersed clay clasts.

The non-fissile diamicton on top of the fissile facies is crudely or distinctly stratified with discontinuous fine sand and silt layers. The textural similarity between the two diamicton facies is striking.

Interpretation

The distinct difference between the fissile diamicton and the non-fissile diamictons indicates important differences of the physical forces, acting on the sediments. The fissility patterns were developed as a response to stresses, acting on the sediments. The distinct boundary between fissile and non-fissile diamictons at site 4 indicates that these stresses were acting on the sediments at a distinct stratigraphic level. The same type of distinct boundary was also found at site 1. The two principally different types of fissures are the low-angle fissures and the vertical fissures. The low-angle fissility is interpreted as thrust faults. The response to shearing stresses has only partly resulted in strain, and the shear stresses have also been released by the development of shear thrusts and a bulk transport along shear planes. The vertical strike-slip faults were developed with horizontal maximum and minimum stresses, i.e. under tensile stress.

A fissility pattern in diamictons has frequently been described and is one of the criteria which are used to recognize lodgement till (Commission on Genesis and Lithology of Quaternary Deposits, Work group 1, circular no. 18). A low-angle fissility pattern is



Fig. 71. Columnar joints in facies E2f at site 4.

characteristic of shearing with low ratios of vertical stress to horizontal stress, and has been taken as evidence of a thin over-riding ice (Broster et al. 1979).

The fissility pattern is formed beneath the moving glacier, and the shearing may affect any subglacial deformable sediment (McClintock and Dreimanis 1964, Boulton 1979, 1982, Boulton et al. 1974).

The striking similarity to the non-fissile diamicton at the top of the fissile facies indicates a genesis similar to these sediments. The distinctly or crudely stratified diamicton is interpreted as proglacial gravity flow sediments (see: The stratified diamicton facies), and the fissile diamicton at site 4 is regarded as subglacially sheared proglacial sediments. The fissility pattern at site 12 is developed in a homogeneous diamicton. The homogeneity was the result of an efficient mixing of material. This may be caused by recycling of debris during repeated deposition and erosion processes at the base of a temperate glacier. It may also be caused by an intense internal strain deformation of subglacial sediments. This is obtained during reduction of the intergranular friction when the pore-water pressure in the sediments is increased. The homogeneous diamicton can thus be interpreted as either debris, released directly from the glacier, or sediments, mixed by intergranular strain underneath the glacier.

The stratified diamicton facies

Sedimentary characteristics

The stratified diamicton facies was investigated in two exposures at site 1. The strata in the two exposures represent different types of stratification, and the two types are about equally represented in the area.

The first type was studied in two sequences, 8 m apart, in the northern portion of the cliff section (1:100-115, Fig. 6). The strata are affected by glaciotectonism and dip slightly towards the east. The silt and clay laminae in the sequence indicate that the bedding-planes in primary position were close to horizontal, and the strata in the logs are adjusted to a horizontal position. The strata (Fig. 72) represent a sequence of approximately 10 m in the upper part of the Alabodarna member. 5 m of strata were hidden behind scree masses below the excavated sequence, and the stratified facies outcropped again in a small exposure in the basal part of the cliff. A fissile crudely stratified facies was also exposed in the basal part.

The stratified facies consists of 10 to 100 cm thick diamicton beds, alternated by 0.5 to 1.2 cm thick reddish brown clay laminae and 0.5 to 100 cm thick beds of laminated silt and very fine sand. Extensive beds of laminated sediments in primary position are rare in the lower part of the sequence. Disrupted and reworked clay clasts are, however, frequent in the diamicton beds. The absence of intrabeds and the presence of clay clasts in the diamicton layers can be connected to the appearance of the lower boundary of the diamicton layers.

Three different types of lower contacts were recognized in the two sequences. Fig. 73a shows an undisturbed clay lamina in primary position overlaid by a diamicton bed. Fig. 73b shows a disrupted clay lamina with the broken pieces clustered. Fig. 73c shows a completely brecciated clay lamina in a diamicton matrix. The figures illustrate a gradual

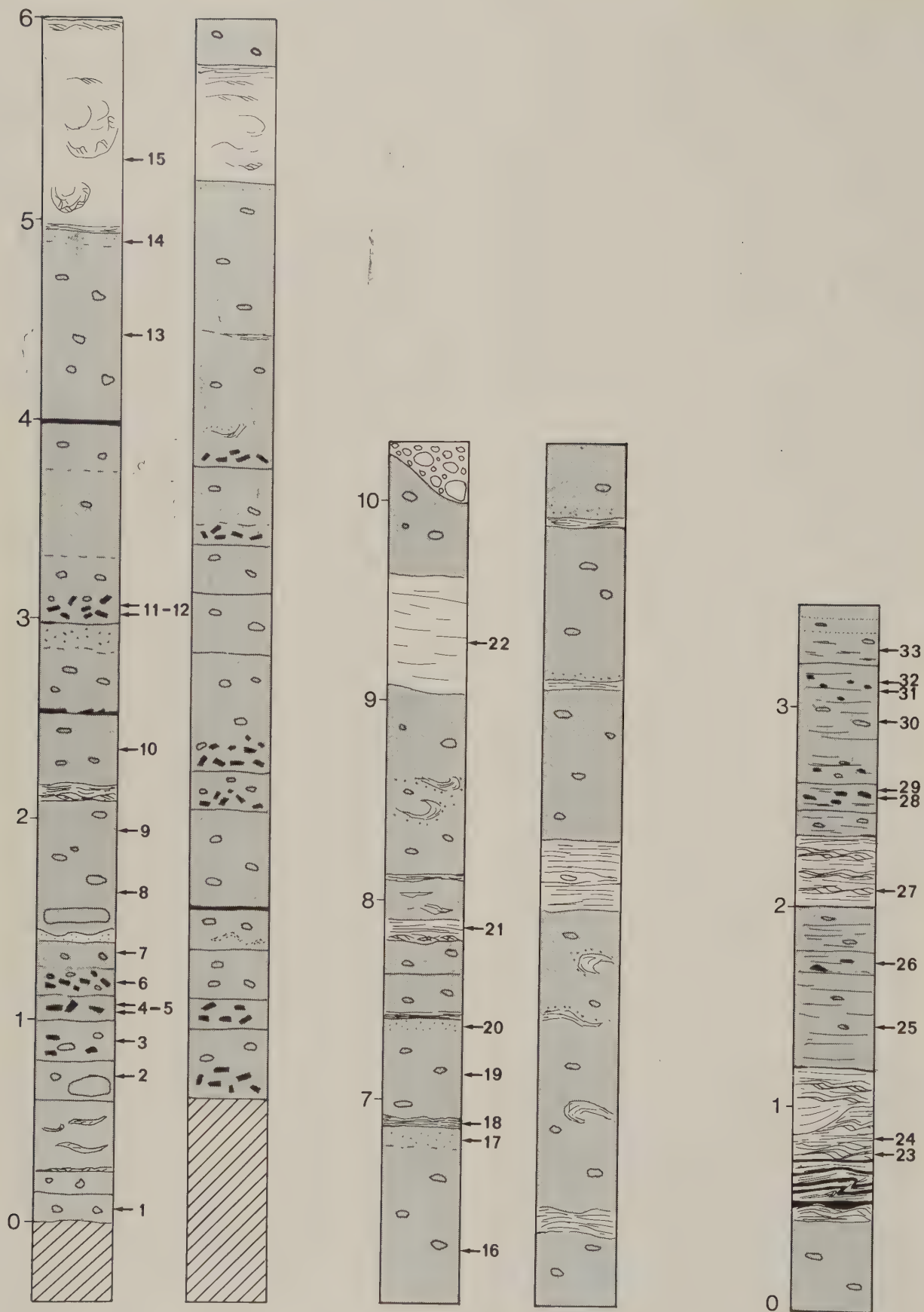


Fig. 72. Logs of sedimentary successions in the stratified diamicton facies at site 1. The two logs, log 1a and 1b, are measured at 1:100-115, 8 m apart. Log 2 is from the exposure at 1:475-485. The numbers to the right of the logs are sample numbers. Grain size parameters of the samples are displayed in Table 6. Intraclasts in black, rock clasts open.

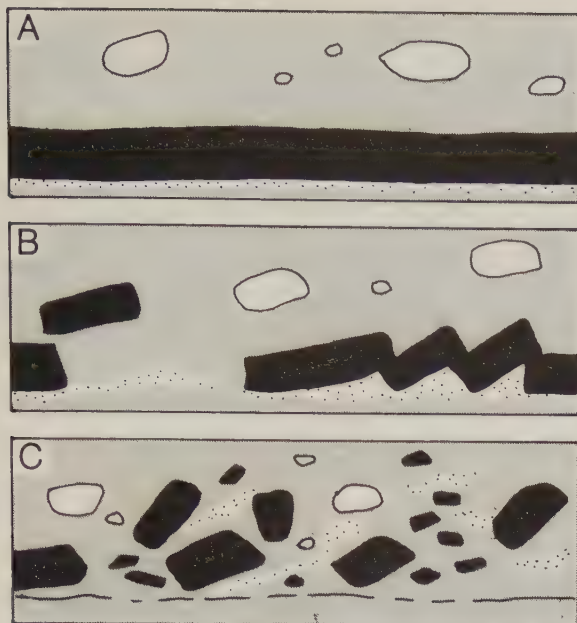


Fig. 73. Three types of lower contacts of diamicton beds in facies E3D at site 1 (section 1:100-115).
a. an undisturbed clay lamina in primary position.
b. a disrupted and clustered clay lamina.
c. a brecciated and incorporated clay lamina.

increase of mixing of adjacent beds. The mixing is obvious when the beds differ in texture and sedimentary structures. If the differences in sedimentary properties are small, the boundaries are hardly noticable, and it is possible that some of the diamicton beds in the logs (Fig. 72) consist of multiple sedimentary units with indistinct boundaries.

The boundaries between the beds in the upper part are mostly more distinct, and beds of laminated silt and clay are generally in a primary position which accentuates the stratification. The diamicton- or silt-supported clay clasts were not to be found in the upper part, and the lower contacts of the diamicton beds are of the first and second type (Figs 73a and b).

The geometry of the beds was difficult to determine due to the narrow excavation. Most of the layers had an equal thickness within the 2.5 m wide excavation. Steeply declining wedge-formed snouts were observed in some of the beds.

The texture of the diamicton beds varies between a homogeneous poorly sorted matrix with dispersed rock clasts (Fig. 74a and b), to a heterogeneous sediment with irregular lenses of fine sand and silt and dispersed intraclasts of clay (Fig. 74c and d). The clay clasts range in size from a few millimetres to 8-10 cm. The rock clasts are generally less than 5 cm in diameter, but clasts up to 32/25/5 cm were found. Several of the beds in the lower part of the sequences can be divided into a lower part with clay intraclasts and an upper part with rock clasts. The clay clasts are absent in the upper part of the sequence. The beds in this part can often be divided into a lower homogeneous diamicton with rock clasts and an upper zone without clasts. A slight enrichment of granules and pebbles was noticed in the boundary between the two zones.

Differences in grain size composition be-

tween the two zones are obvious. The top samples always show a marked decrease in 7 to 11 ϕ size grades, compared to samples from the lower part, and increased values of 3 to 5 ϕ size grades.

The stratified diamicton facies at 1:475-485 differs from the diamicton at 1:100-115. A sequence with 3.5 m of diamictons interbedded by fine sand, silt and clay was excavated (Fig. 72, log 2). The stratification of the interbeds is slightly disturbed. The primary sedimentary structures consist of A-type and B-type ripples superimposed by laminated silt and, in the lower part, also by clay.

The diamicton can be split up into 2 to 30 cm thick beds separated by very thin, fine sand laminae. The individual layers are continuous and of almost uniform thickness within the exposure. The structures within the individual beds is a crude stratification or lamination (Fig. 74e).

Tabell 6. Grain size parameters of samples from the stratified diamicton facies at site 1. Sampling spots are shown in Fig. 72.

Type of bedding	Md ϕ	Mz	ϕ	ϕ	Sample No.
Homogeneous diamicton beds with rock clasts					
	5.3	6.6	2.7	0.49	1
	5.3	6.4	2.5	0.45	2
	5.4	6.4	2.4	0.44	7
	5.1	6.1	2.4	0.42	8
	5.5	6.8	2.7	0.45	9
	5.5	6.6	2.6	0.44	10
Diamicton beds with inversed grading on the top					
lower	5.2	6.2	2.3	0.44	13
upper	4.5	5.8	1.6	0.57	14
lower	5.3	6.5	2.6	0.42	16
upper	4.9	5.8	2.1	0.46	17
lower	4.9	6.2	2.9	0.56	19
upper	4.7	5.9	1.9	0.54	20
Diamicton beds with rock clasts and intraclasts					
	6.3				3
	5.7				6
Diamicton beds with intraclasts					
matrix	8.6				4
i.clasts	11				5
matrix	7.4				11
i.clasts	9.5				12
Crudely stratified diamicton with rock clasts					
	4.8	5.9	1.9	0.51	25
	4.7	5.8	1.8	0.50	30
	4.3	5.5	1.7	0.54	33
Crudely stratified diamicton with intraclasts					
matrix	5.3	6.4	2.3	0.44	26
matrix	9.4				28
i.clasts	11				29
matrix	6.1				31
i.clasts	9.6				32
Cross-lamination and disturbed cross-lamination					
	4.6	4.7	1.0	0.37	15
	3.7	3.8	0.7	0.28	23
	4.6	4.7	0.81	0.26	27
Parallel-lamination					
	5.4	5.5	1.2	0.33	18
	4.5	4.6	1.3	0.44	21
	4.3	4.4	1.4	0.28	24
Crudely laminated silt					
	5.2	5.3	1.2	0.32	22

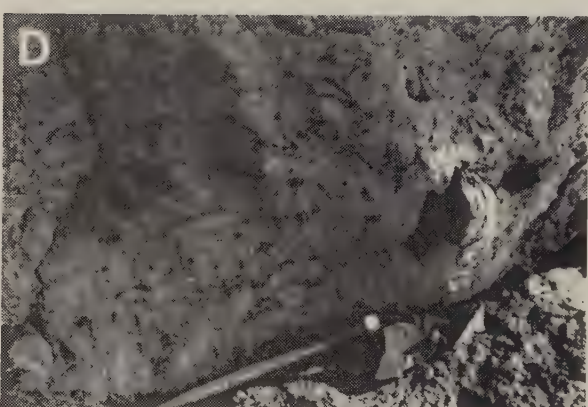
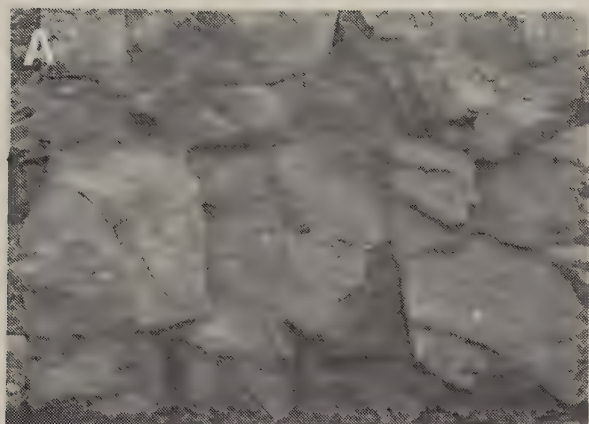


Fig. 74. Diamicton beds from the stratified diamicton facies. A and B. Homogeneous beds. Rock clasts are dispersed in the matrix. C. The boundary between a lower bed with rock clasts and an upper bed with intraclasts. C. A diamicton bed with silty clay intraclasts in a silty matrix. E. Internal, crude stratification in a diamicton layer with clay intraclasts.

The textural composition of the diamicton layers is of two types. One type has a grain size composition of the matrix which is slightly coarser than the average composition in the homogeneous beds from 1:100-115. The mean grain size (Table 6) is between 5.5 and 5.9 ϕ compared to 6.1 to 6.8 ϕ in the homogeneous beds. The diamicton contains scattered rock clasts. The clasts are blade- or disc-formed and lie with their a/b-planes parallel to the bedding planes. The clasts are rare, and elongated clasts are still rarer. The number of measured elongated clasts never exceeded 8 in an individual layer, and the readings have no statistical significance.

The second type of grain size composition has a mean grain size, exceeding the average mean of the homogeneous beds. The individual layers have, however, an internal structure similar to the first type with streaks or a diffuse pseudo-stratification in different colour shades (Fig. 74e). Rock clasts are extremely rare or absent, and the main part of the clasts are intraclasts of clay. The clay intraclasts are randomly dispersed.

Interpretation

The stratified diamicton facies of the Ålabodarna member shows three main types of sediment, diamicton beds, clay laminae and parallel- and cross-laminated fine sand and silt.

The clay laminae have a very high clay content and were deposited by fall-out from suspension in a still-water basin.

Current ripples in the laminated fine sand and silt indicate that low velocity underflows brought sand and silt into the place. The combination of fine sand ripples, coated by draped and parallel lamination, is interpreted as C and D zones of the Bouma sequence (Bouma 1962). The turbidites were low-concentration flows with a well developed bed load (Nilsen, Walker & Normark 1980).

The lower contact of the diamicton beds is often erosional. Slump structures in the substratum, crumpled up clay laminae, and incorporation of the substratum are evidences of shearing along the bed. This, in combination with the poorly sorted grain size composition of the diamicton layers, limits the interpretation to two possible processes, debris flow (including till flow) and till, deposited beneath a moving glacier. Important criteria for lodging processes and subglacial shearing are missing, and the repeated alternation of still-water sedimentation and sedimentation of the diamictons also makes the subglacial processes less likely. A gravity-induced movement of a sediment-water mixture is more likely. The matrix strength was high, and clasts were suspended in the flow. The high silt and clay content increased the viscosity of the matrix-water mixture. The diamicton beds in the upper part of the sequence at 1:100-115 show a faint inverse grading with decreasing silt and clay content in the uppermost part. This is an often observed condition in debris flows (Fisher 1971) and is due to fluidisation and water interchange with the overlying water.

Subaqueous mudflows often show erosional lower contacts, sometimes with incorporation of lower sediments (Middleton and Hampton 1973, 1976). The shear strength at the lower contact of the clay laminae in the substratum was reduced by an increase of the porewater pressure underneath the

clay from the loading of the debris flow, and the clay lamina was moved and incorporated in the flow. Sometimes the lower sediments were only slightly sheared without intercorporation, and in a few cases a consolidated surface of clay was unaltered.

The crudely stratified layers at 1:475-485 consist of a combination of clay clasts, streaked-out lenses of silt and a poorly sorted matrix. These layers are interpreted as a mixture of silt and clay sediments eroded from the substratum and mixed with diamicton debris from the original flow. The pseudostratification of the beds was produced by internal shearing of the flow. The incorporation and mixing of the sediments took place in an earlier stage of the flow, and the lower contacts at 1:475-485 are non-erosive. The low ratio of rockclasts is probably due to a slight reduction in matrix strength, compared to the more proximal facies at 1:100-115. The few clasts in the crudely stratified facies are all blade- or disc-formed and did not settle in the high-viscose laminar flow.

The distinct boundary between the individual beds is interpreted as the boundary between the different debris flows. The thin silt laminae, separating the individual beds, were deposited from turbidity flows on top of the slower moving high density debris flows.

6.1.2 Environmental interpretation

The 15 to 20 m thick strata that were deposited presuppose an intense debris input into the basin. The sediment gravity flows were only responsible for a redeposition and a spreading of the sediments in the basin.

The rock fragments and the reworked marine sediments were transported from the Kattégatt area into the Ven-Grumslöv area. The transport of clasts in a low gradient area and even with a slight upslope inclination can only be made by glacier ice. Striated clasts are also present in the sediments, and this is generally used as a criterion of basal transport in a glacier.

The interpretation of the unload of the debris is more problematic. The debris may have come directly from underneath a grounded glacier or by melting out from floating ice or icebergs. The texture of the sediments with very few clasts is very similar to the texture of glaciomarine sediments, deposited from fall-out from suspension and melt-out of ice-rafted debris. The high silt and clay content in the sediments must not necessarily be interpreted as fall-out from suspension. These size grades were picked up from the floor of the Kattégatt Sea by the glacier, and the grain size composition of the debris in glacial transport was probably dominated by silt and clay. The depth of the water is essential for the interpretation. A floating ice or debris-laden icebergs demand a water depth of more than 9/10 of the ice thickness. The water depth cannot directly be interpreted from the sediments. The strata on top of the Ålabodarna member indicate, however, a rapid, but not dramatic transition into subaerial conditions, and this may be an indication of a rather small water depth in the basin during the deposition of the Ålabodarna member. The absence of ice-rafted clasts in the clay laminae and in the varved sediments on top of the diamictons indicates that floating icebergs were rare in the basin.

The similarity in textural composition within the area implies an intense mixing

of the debris during the glacial transport which only can be expected at the base of a wet-based glacier. Evidences of lodging processes are, however, extremely rare. The fissile sediments at site 4 were sheared, but not lodged. The only sediment that may be deposited directly from the glacier is the fissile, homogeneous diamicton at site 12 in the mainland.

The thick strata of diamictons thus indicate a high amount of basal transport but reduced basal deposition.

Lodgement is favoured by increased ratio of normal forces relative to shear forces (Boulton 1974). The normal forces are increased by increase in ice thickness. The effect of basal water pressure or the reduced weight of the ice, when being in a water body, reduces the normal forces. The debris in the basal part of the glacier was continuously released by basal melting. The load of the immersed glacier was low while the forward directed forces from the moving glacier were high. The basal meltwater did not easily escape from the immersed base. A high water content, the pore water pressure and the silt and clay-rich matrix favoured a subglacial flowage of the debris. The water-soaked debris was moved beneath the glacier and finally squeezed out from under the glacier margin. Such processes were interpreted by Okka (1955) and Price (1969, 1970) as responsible for the formation of marginal moraines of De Geer type.

When reaching the glacier margin, the debris was slowly moved forwards by gravity forces along a proglacial slope.

The fissility in the lower part of the diamicton sequence at site 4 was then formed during a glacial advance over proglacial till-flow sediments. The fissile diamicton at site 12 may be of the same origin but with a more intense internal strain deformation and thus a higher degree of mixing. It may also be a real basal deposition. The geographic position of the sediments during formation was probably in the lower part of the bedrock elevation, and it is probable that the base of the glacier was close to the proglacial lake level. In that case the reduction of normal stresses was less than in the deeper Ven area, and conditions for lodging were more favoured.

The Ålabodarna member is the lowest stratigraphic unit exposed in the cliffs and in the clay pits. The lower boundary was never reached in the sections, and the transition from the Uranienborg member to the Ålabodarna member was only recorded from the boring on Ven. The vertical distribution of sediments on the sea floor to the north of the Kyrkbacken village can be compared to the vertical distribution of strata in the boring (site 17). The sea floor sediments consist of a diamicton at approximately -30 m to -10 m (Hörnsten 1979). The diamicton has the same vertical position as the Ålabodarna member in the boring, and the thickness, about 20 m, corresponds to the maximum thickness measured in the sections. The diamicton on the sea floor is probably strata from the Ålabodarna member in primary position, and the vertical distribution on the sea floor corresponds to the thickness of the member. The boring on Ven. The vertical distribution of sediments on the sea floor to the north of the Kyrkbacken village can be compared to the vertical distribution of strata in the boring (Fig. 1). The accordance in vertical position of a thick diamicton unit on top of sorted sediments indicates that the Ålabodarna member has a primary vertical position

at approximately -30 to -10 m and that the maximum thickness, 21 m, of the Ålabodarna member, measured in the sections, approximately corresponds to the real thickness of the member.

The environmental interpretation of the Ålabodarna member is that the whole sequence, except for the lower fissile facies, was deposited in a subaqueous proglacial environment. The lake was not deep enough to allow the ice to float or calve, but the lake was deep enough to allow the sedimentation of clay laminae. A very rough estimation gives a lake level about equal to the present sea level in the area.

6.2 Glumslöv member: a braided river plain and a glacial lake environment.

The strata from the Glumslöv member are by volume the most important in the Ven-Glumslöv area. Borings from the hill-row at Glumslöv-Rönneberga indicate that the silty and sandy sediments are important constituents in the hills (Adriellsson et al. 1981), and most of the investigated sections are dominated by these sediments.

The Glumslöv member is the uppermost strata, affected by Möllebäcken glaciotectionics, and the upper boundary of the member was defined by the unconformity that separates the member from younger sediments (see chapter 5.3). The deformation is a disadvantage in the palaeo-environmental reconstruction, as the dislocations imply displacement from the primary position, and the surface of angular unconformity represents an erosive phase with removal of unknown quantities of strata.

The accessibility of the sediments in some of the cliffs permits, however, close investigations of proportionally long sequences, which give good possibilities for stratigraphic correlations and environmental interpretations.

The lithostratigraphic sequences show repeated changes between sand, silt and clay units, pointing at a complex environmental history with changes in depositional conditions and energy.

6.2.1 Sedimentary characteristics and interpretations

Four sedimentary successions from different parts of the area were chosen in order to show the changes in the sedimentary environment during the Glumslöv interphase.

Log 1

Log 1 (Fig. 75) represents an adjusted vertical sequence of the dislocated units 1a:3 to 1a:8 at site 1:125-225. A brief description is given in chapter 5.1.1.

Sedimentary characteristics

The lowermost unit, unit 1a:3, is a discontinuous bed of gravel, which rests on a scoured surface on top of the Ålabodarna member. One channel trough, approximately 0.5 m deep and 2-3 m wide, was eroded into the lower member. The trough is infilled by

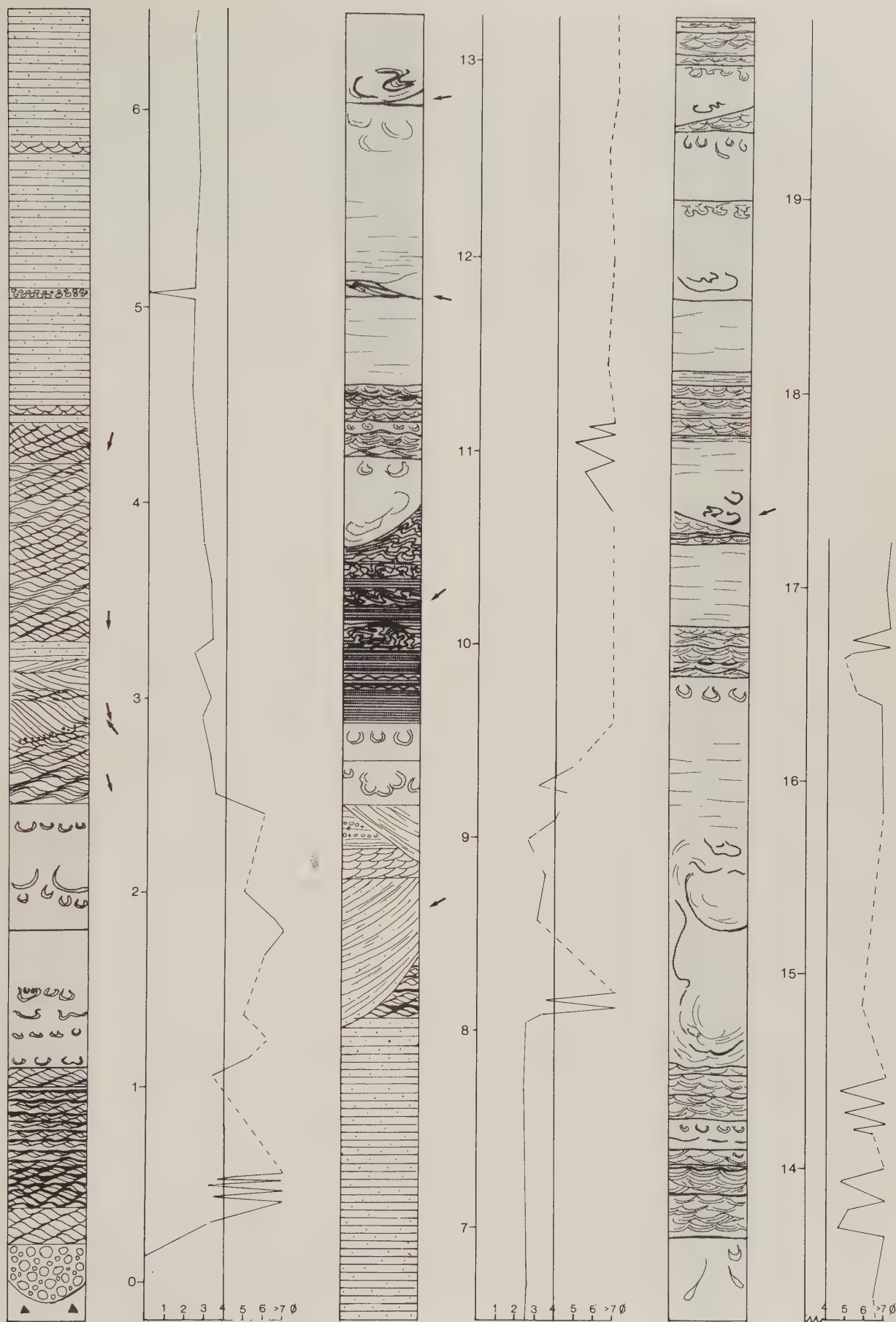


Fig. 75. Log 1, comprising units 6, 7, 8, 10, 11 and 12 (Fig. 62) at 1:125-225.

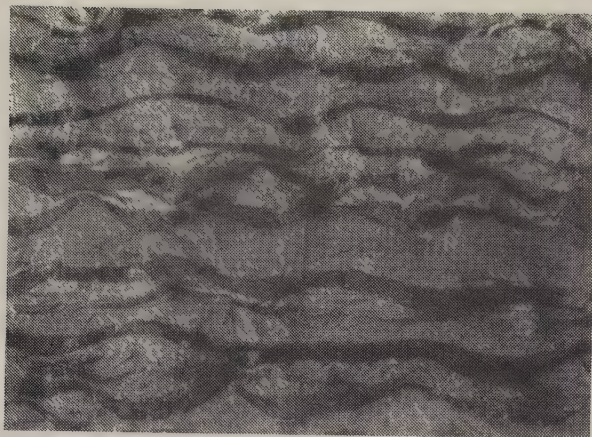


Fig. 76. Alternating bed facies of unit 1a:4. The clay laminae have distinct upper and lower boundaries. The individual sand layers consist of single ripple sets.

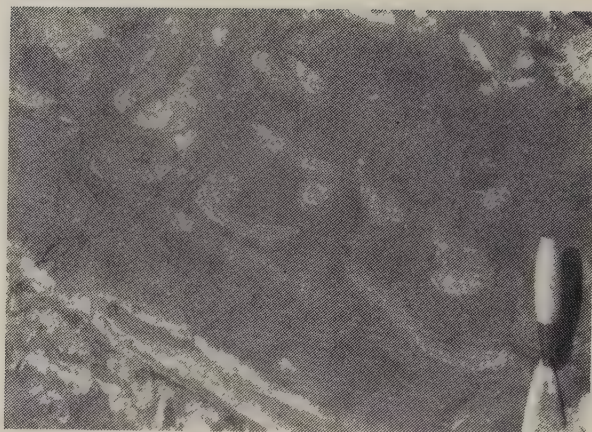


Fig. 77. Diffuse layers of homogeneous silt with load cast structures.

poorly-sorted sandy gravel. A discontinuous sheet of gravel lies on top of the substratum outside the channel. The bedding is structureless or a poorly defined horizontal bedding.

The alternating bed facies (unit 1a:4) consists of distinct clay and fine sand layers which form couplets (Fig. 76). Such alternations of clay and fine sand or silt are generally called a varved sediment (De Geer 1912). The sequence contains 36-40 varves.

Both the sand/clay contacts and the clay/sand contacts are fairly sharp, and erosional contacts between a clay lamina and a sand layer occur. Gradational transitions are absent.

The thickness of the individual sand beds ranges between single lenses of ripple sets to continuous layers up to 5 cm thick. The thicker layers sometimes, but not always, consist of several ripple sets, separated by thin draped laminae of silt. The sedimentary structures in the sand layers are dominated by A-type ripple lamination, generally with high angles of climb in the range 30° to 40°.

The clay laminae range in thickness from 0.5 cm to 1.2 cm, and the laminae are composed of a dark red upper part and a slightly lighter brownish gray lower part. The transition between the two parts is gradual and forms a diffuse zone of less than 0.1 cm thickness.

Unit 1a:5 consists of two thick beds, 65 and 70 cm thick, with a gray homogeneous silty matrix with convoluted balls and pillows of laminated silt in diffuse parallel layers (Fig. 77).

Fine to very fine sand makes up the 6 m thick sand unit (unit 1a:6). The mean grain sizes show small variations and range between 2.2 Ø and 3.5 Ø. The sand is well to moderately sorted.

Planar lamination (Facies B2) and cross-lamination (Facies B3) are exceedingly predominating, and their total percentage share amounts to more than 80 %. Cross-bedded sand (Facies B1) is concentrated to the lower part of the sand sequence where ripple cross-lamination and large scale foresets are intimately connected (Fig. 78). The complex interwoven coset is covered by a 12 cm thick tabular set of cross-bedding with planar lamination on the top. A small wedge-shaped fissure, 0.8 cm wide and 20 cm deep, filled with structureless sand was developed from the top of the planar laminated sand. Such small wedges are common at this stratigraphic level along the entire section.

The planar lamination in the upper part of the sand contains parting lineations. A few stringers of cross-laminated sand (type A ripples) and a couple of coarse sand veneers are interbedded in the planar laminated sand.

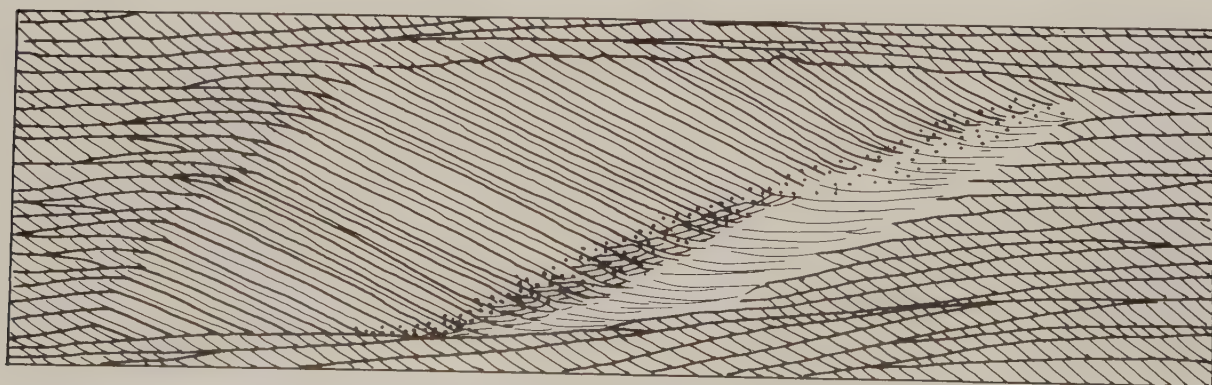


Fig. 78. A microdelta or solitary bar, developed from the growth of a ripple foreset. The foreset laminae in the middle part of the picture have angular basal contacts with regressive ripples at the toe. The foreset laminae in the right part abut the floor with slightly tangential contacts. The zone of flow impinge in front of the foreset slope has a diffuse horizontal lamination.

The planar laminated sand sets are very consistent laterally and were traced to the limit of the excavation, which means for at least 50 m, without changing character. No significant scoured surfaces or channel erosions were observed not even in connection to coarse-grained veneers.

The grain size distribution of samples from the upper 5 m of the sand unit is very uniform, with mean grain sizes between 2.23 and 2.5 ϕ . The sand is well sorted with 80% of the samples with a standard deviation between 0.35 and 0.5. Skewness (Sk_I) is fineskewed or nearly symmetrical (ranges between 0.31 and -0.06). Ripple samples are generally more fineskewed than planar laminated samples. The exceptional coarse sand veneers are poorly sorted and fine skewed.

The sand grades into 30 to 40 cm of alternating bed facies with parallel-laminated silt and cross-laminated fine sand to coarse silt. This unit is cut by a series of erosional channels. Channel geometry varies between narrow channels with semicircular cross-profiles and broad shallow channels with curved sides and flat bottoms. The channel infillings are of different types from well to moderately sorted sand with layers, confirming roughly to the channel shape, to poorly sorted sand with dispersed pebbles or poorly sorted sandy silt with convolute bedding. Ripple lamination and planar lamination occur on top of some of the channels. This heterogeneous unit with erosional channel infillings can be traced as a 1 to 2 m thick zone along the entire cliff-cutting.

Silt and clay facies (unit 1a:8) completes the sequence. Grain sizes larger than coarse silt are almost absent in the 10.5 m thick unit. In the analysed samples a maximum of 2 to 3 % very fine sand has been recorded.

The silt and clay unit consists of clay laminae and cross-laminated, parallel-laminated or homogeneous silt. Erosional and different types of deformation structures are frequent throughout the whole sequence. A distinction between primary sedimentary structures and deformation structures is not always readily made, as the deformation is often intimately connected to redeposition after desolution and decomposition. Roughly estimated, more than 50 % of the strata is penecontemporaneously disturbed, distorted or deformed.

The lower 1.5 m of the unit are dominated by clay laminae, ranging in thickness from 0.1 to 2.5 cm (Fig. 79a). The silt occurs as thin streaks, graded-bedding laminae, parallel-laminated layers or cross-laminated layers with an apparently random vertical distribution. Individual varves or varve couplets with distinct summer and winter layers cannot readily be distinguished. Upper and lower boundaries between the clay laminae and the silt laminae are either distinct, diffuse, graded or erosional. In a few cases the contact between a clay and a silt layer is erosional with a subparallel system of small longitudinal scour marks about 0.4 cm wide and 0.2 to 0.3 cm deep (Fig. 79b). The deformation in the clay-dominated part is a plastic deformation with décollement-like small-scale fold structures often with overturning and sometimes formed as intraformational recumbent folds (Fig. 79a). The distorted clay and silt laminae are well-defined, and the individual laminae are not dissolved or decomposed.

At 1.5 m of the silt and clay facies, the silt layers increase in thickness and become consistently thicker than the clay laminae

(Fig. 79c). The clay laminae continue to vary in thickness but with a decreasing amplitude between 0.2 and 1.5 cm.

The thickness of the silt layers varies from thin streaks to more than 1 m thick beds. Varve couplets are not readily distinguished, even if an alternation between clay laminae and silt layers is obvious in some parts of the sequence. Transitions from clay to silt are generally distinct. The contacts are sometimes erosive with a subparallel system of weakly undulating small scour marks (as in the lower part of the unit). Sometimes the clay is eroded in small symmetrical wavy troughs. Transitions from silt to clay are either sharp or graded. The clay laminae are homogeneous or graded. Grading is most evident in the lower part of some of the thick clay laminae. The graded clay laminae occur both when silt to clay transitions are gradual and silt to clay transitions are sharp.

Primary sedimentary structures in the upper part are predominated by cross-lamination. Symmetrical ripples, often sharp crested with rounded troughs, are most frequent (Fig. 79d and e). Vertical form index ranges between 4.9 and 6.6.

Complex ripple sets rarely exceed 10 cm in thickness. The beds of symmetrical ripples are formed as sets with form-discordant boundaries (Fig. 79e) or as sets with ripples in phase (Fig. 79d) often with chevron-like structures (Boersma 1970).

Asymmetrical ripples, parallel-lamination and graded bedding are subordinate.

Other types of primary sedimentary structures are parallel-laminated and draped laminated silt. 1 to 3 cm thick graded beds also occur. The mean grain size in the silty layers ranges between 4.7 and 7.2. The skewness of the grain size distribution shows large variations. The sharp-crested ripples have values between -0.96 and -0.12. The asymmetrical ripples are fine-skewed or strongly fine-skewed (0.24-0.37). Draped lamination and parallel-lamination are mainly strongly fine-skewed (0.24-0.49).

Besides the cross-laminated and parallel-laminated silt-beds with bed thickness generally between 10 and 15 cm, there are thicker beds of reworked sediments. These beds lie either as layers with approximately uniform thickness on undisturbed strata or as lens-formed layers with a lower erosive contact in broad shallow troughs with flat bottoms. The deformed strata range in unit thickness from a couple of centimetres to more than 3 m. The deformed strata were frequently traced laterally into a curved surface of displacement, a slump scar.

The internal structures vary from bottom to top in the deformed beds, and up to four zones can be recognized. The first lowermost zone (zone A) consists of slump structures with folding and thrusting of the stratified sediments (Fig. 80a). Preserved primary sedimentary structures can be recognized, and the lamination of the sediments is only partly dissolved into a more homogeneous silt. Sometimes the laminated sediments constitute isolated slabs, lenses or large chunks in a more homogeneous matrix.

Next zone (zone B) consists of homogeneous silt. The transition from the lower unit is generally rather distinct. The homogeneous part appears structureless. When dried, it splits, however, into numerous horizontal thin laminae (Fig. 80b).

The transition from the homogeneous part to the uppermost part is often diffuse. When distinct, the uppermost zone (zone D)

consists of parallel- or cross-laminated silt (Fig. 80e). A diffuse transition zone (zone C) is, however, more frequent, and often zone D is missing (Fig. 80d). Zone C is convoluted with balls and pillows of laminated silt in a homogeneous matrix. Often individual pouches have sunk into the homogeneous part (Fig. 80c).

A general relation between bed geometry and thickness on one hand and internal structures on the other hand is obvious. Thick lens-formed layers in broad shallow troughs with erosive lower contacts generally consist of A-B-C or A-B-C-D zones. Medium to thin beds with non-erosive lower contacts consist of B-C, B-C-D, D or C-D zones.

Interpretation

The gravel at the base of the member fills up a channel, which most likely was eroded by the current that also carried the gravel as bedload. The channelized current flooded and scoured the surrounding surface and deposited the thin sheet of gravel outside the channel.

The debris flows in the substratum and the strata on top of the gravel were deposited in a subaqueous environment, and there is no reason why the gravel should not have been deposited in the same environment by a submerged underflow current, probably just coming out from underneath the glacier. The proximity to the glacier is interpreted from the substratum.

The poorly sorted debris underneath the glacier is considered the source of the gravel. This is interpreted from the petrographical content (Fig. 7a and Fig. 63). The proportional high amount of gravel-sized clasts makes an enrichment by fluvial processes most likely. The lower diamictons are extremely poor in this grain size, and a deposition from the traction load of a debris flow is not likely.

The varved sediments of the alternating bed facies (unit 1a:4) is the result of two different sedimentary processes and is interpreted as the result of bottom currents from density underflows, alternated by fall-out from suspension (e.g. De Geer 1912, 1940, Kuenan 1951, Mathews 1956, Ashley 1972, Gustavson 1975). The sand deposition with current bedding and erosional contacts to the lower clay laminae was caused by fluid shear beneath turbulent currents in a density underflow. The cross-laminated sand was produced by short-lived flow events, and each flow was represented by a single ripple set with only one stratification type. Some of the sand layers were formed by one single flow event, and some were formed by multiple flow events with sand sets separated by erosion or by draped silt laminae. It is shown by Gilbert (1972) that turbidity underflows in a proglacial lake is not a continuous event through the melting season. Underflows may in some years occur only once or twice or even not at all, while in other years significant underflows may occur during as much as half of the melting season. The turbidity currents in proglacial lakes are normally generated by turbid meltwaters during periods of high meltwater discharge (Gustavsson et al. 1975, Ashley 1975). The coarser sandy or silty "summer"-layers are often abruptly covered by the clay laminae (Agterberg and Banerjee 1969, Ashley 1975).

Walker (1963) suggests that the rapidly deposited sand in proximal environments

might be "abandoned" by the main turbidity current and hence never grades upwards into finer grain sizes.

The clay laminae are deposited by fall-out from suspension. The dividing into two parts of each lamina may be the result of seasonal variations in the glacial lake. During the summer season the glacial lake water was agitated by the influx of meltwater into the basin and by wind action on the lake surface, and the lower part of a lamina was settled. During the winter season the glacial lake surface was covered by winter ice, and the influx of meltwater ceased, and the finest colloidal clay particles could settle.

The two thick units with zones of convoluted lamination are interpreted as formed by turbidity currents. Each diffuse zone is regarded as one layer. The sedimentation rate was high, and so was the pore water excess. The convoluted zones appear to have undergone simultaneous deformation and thereafter have been eroded by a new turbidity current, which in its waning stage deposited the overlying silt bed.

Each diffuse layer is thus the result of one single high-density turbidity current, and the repeated pattern indicates a non-uniform current flow. The mixing between the different beds shows that the sediments were not stabilized before a new underflow current entered the place. The distinct clay lamina between the two units indicates a period of slow sedimentation rate with fall-out from suspension between the periods of repeated pulsatory underflows.

The interplay between the different structures in the complex sand body (Fig. 78) on top of the convoluted layers forms a perfect model of the growth of a small delta or micro-delta (terminology after Potter and Pettijohn 1963). The type A ripples in the upstream part were deposited from traction load on a current ripple bedform in the lower flow regime. The continuity in upstream direction shows a uniform, steady flow. A sudden change in flow conditions at a certain point made a ripple grow and grade into a foreset slope. The production in tractive stress may be the result of an expanding flow either by lateral widening of the channel or by a backwater effect, or both. The backwater effect is achieved when a stream flows into a basin or a lake and the limit of backwater coincides with the theoretical upstream limit of topset aggradation (Jopling 1963, 1965).

The delta grew and split up into topset, foreset and bottomset components of a small idealized delta body of Gilbert type.

In the middle part of Fig. 79 the foreset slope has grown close to a state of equilibrium, and the delta front continues to grow laterally. As the delta advances, the limit of backwater retreats further upstreams (Lane 1955), and topset aggradation continues. The flow and the particle trajectory over the delta front follow the paths, outlined by Jopling (1961, 1962), and the structures can be explained by hydraulic phenomena in the vicinity of the delta front with an inertia-dominated jet with flow separation, free-turbulence shear flow, zone of zero velocity (zone of flow impinge), zone of backflow and zone of coflow (Jopling 1961, 1963, Boersma et al. 1968, Wright 1977).

The foresets in the central part of the delta are planar and abut the floor of the basin with an angular contact. This is favoured by low flow velocities (Jopling 1965b). The transition into a slightly tangent-

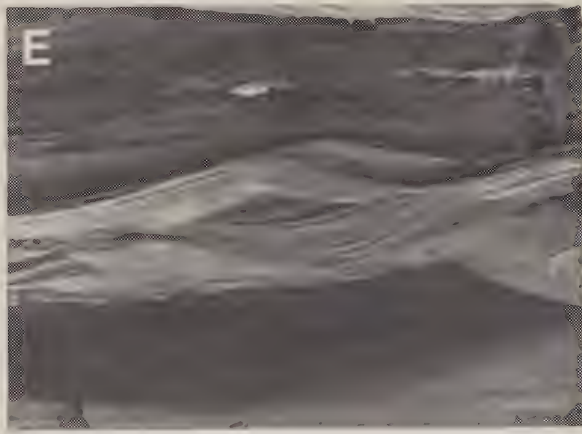
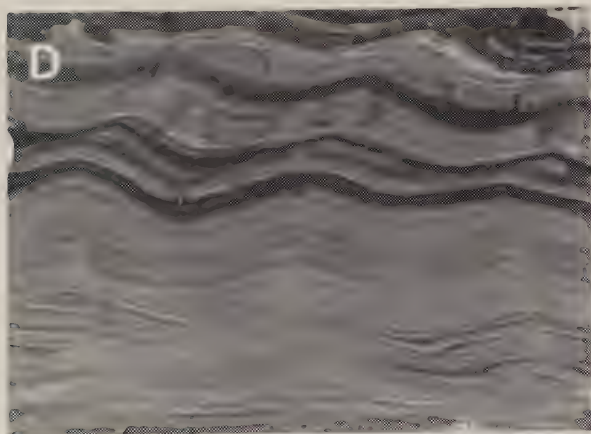


Fig. 79. Silt and clay facies, unit 1a:8. a. The lower part of the unit with clay laminae thicker than silt laminae. Note the eroded clay surface in the middle part and the fold structures in the upper part. b. A clay surface with sub-parallel weakly undulating erosional marks. c. The upper part, dominated by silt layers. Mark the widely spaced clay laminae. d. Sharp-crested ripples in phase. e. Form discordant sets of sharp-crested ripples.

ial contact (at decreased depth of the foreset to about 20 cm) is the result of an increasing sedimentation of sediment load suspension on the toeset and bottomset and can be interpreted both as a result of increasing flow velocity or as a decrease in the depth of the basin (Jopling 1965b). The change of the depth of the basin is obvious with a gradual decrease from 80 cm to zero, and there is no change in ripple structures of the topset bed, indicating any change in flow velocity of the streamflow, which leaves the topset and by-passes the delta front. The transition into slightly tangential contact shows a transition from low depth-ratio to moderate or high depth-ratio in the basin.

A low water depth can also be interpreted from the planar-laminated sand above the interwoven set. Plane bedforms of upper flow regime are developed beneath flows of high velocity or low depth (Simons and Richardson 1961, Simons et al. 1965).

The emerged fluvial environment is finally confirmed by the appearance of ice-wedge casts, which were developed from a drained surface in a subaerial environment.

The lower part of the Glumslöv member can thus be interpreted as a gradually increase of the fluvial influence, which started as density underflows in a proglacial lake, continued as a 'Gilbert-type' distributary mouth bar (Wright 1977) and finally ended up in an emerged subaerial fluvial environment. The stream came from the north.

This thick sequence of planar lamination in the upper part of the sand unit is of medium to fine sand and was deposited in the upper flow regime. The consistent lateral extent and the absence of cross-bedding and scoured surfaces in the 3.7 m thick sequence of planar lamination give an indication of a non-channelized flow, and the most probable origin of the extensive flat-bedded sets is sheet flood deposits. Sheet flood deposits are made by flood flows, spreading out from a channel on to a plain. The flow is shallow, and mostly upper flow regime conditions prevail (Rahn 1967). Flood deposits at Bijou Creek, Colorado, were investigated by McKee et al. (1965). During one single major flood event a maximum of about 3.5 m was deposited. Estimations indicated that 90 to 95 percent of all deposits consisted of planar laminated sand.

The planar laminated sand in the section is connected to fluvial sand in the bottom and gradually grades upwards into silt and clay facies. The flooding was probably not the result of an ephemeral stream but rather the result of a more permanent elevation of the water level.

The sediments on top of the planar-laminated sand consist of an alternation of cross-laminated sand and clay laminae. The ripples were formed by currents, and the sedimentation from traction load was intermittently alternated by a slow fall-out of mud from suspension.

The trough-formed infillings in the upper part of the alternating bed facies show different types of bedding. Structureless sand and silty sand were rapidly deposited from debris flows or from high-concentration turbidity flows. The upward change into ripple cross-lamination in some of the infillings can be connected to the Bouma sequence.

The sedimentation of the lower part of the silt and clay facies (unit 1a:8) was a slow settle of clay particles, which generally were the only grain size, available

for deposition. Now and then turbidity underflows brought silt into the calm environment. The settling of clay is a process that continues all the year round. In a proglacial lake it is only interrupted by short periods when turbidity currents are passing (Gustavsson 1975, Ashley 1975).

Sets of cross-laminated silt with type A and type B ripples were deposited by settling during fluid shear at the base of the turbulent currents. Silt streaks or coarse silt partings, which sometimes swell into ripple embayments, were also formed in lower flow regime with low applied shear stresses. Under such conditions the traction carpet is restricted to rolling and saltation in a zone, only a few grains thick (Walker 1967). A turbidity current of short duration may leave behind a thin lamina or a thin ripple set, deposited from the base of the traction carpet.

Thin graded silt layers with a diffuse transition into a clay lamina are the result of sedimentation from the suspension cloud in the tail of a bypassing current or in the last waning stage of a turbidity current. The incomplete Bouma sequence, A and B parts are missing, and the high ratio of clay relative to the C and D parts, indicate low-concentration turbidity flows (Nielsen et al. 1980).

Erosional marks, similar to the sub-parallel grooves on some of the clay surfaces, were formed experimentally by Allen (1969). He found that the recti-linear or meandering small grooves were formed parallel to the flow and at low flow velocities. Their origin and spacing appeared to depend on small sub-parallel stream-streaks of dense suspension in the lower parts of the turbulent boundary layer. An alternation of high velocity and low velocity streaks is a characteristic pattern in the laminar sublayer of subcritical flows (Kline et al. 1967, Corino and Brodkey 1969). The distorted or contorted lamination may be formed by shearing action on the sediments from an overriding force, for instance a turbidity current. It may also be the result of a slow mass movement of saturated sediments on an inclined slope (Coleman and Gagliano 1965).

The difficulty to find a repeated pattern with silt and clay laminae, forming varve couplets, indicates that the lake bottom was probably not affected by cyclic inputs of suspended matter from underflow currents. Two questions arise: What was the source of the turbidity currents? Why are seasonal variations not obvious, as they generally are in the glacial lake environment? Low density turbidity currents can be generated by (1) the infusion of a muddy stream into the lake, (2) the dilute tail of a turbidity current with a higher density, (3) storm waves stirring up sediments in more shallow parts of the basin (Moor 1969, Rupke and Stanley 1974). Muddy streams, running into glacial lakes, are connected to periods of high meltwater discharge. This is strongly related to seasonal variations. Storm waves in a glacial lake are also connected to the short summer season when the lake is unchained from the winter ice. High density turbidity currents, gradually grading into low density currents, are most probably related to sliding and slumping of delta or lake sediments. This can be a seasonally controlled process (Shaw 1977), occurring during draw-down of lake levels at the end of the ablation season. It can also be the result of fluidization and tixotropy of silty glacial lake sediments, caused by the mechanical impact or shock as the

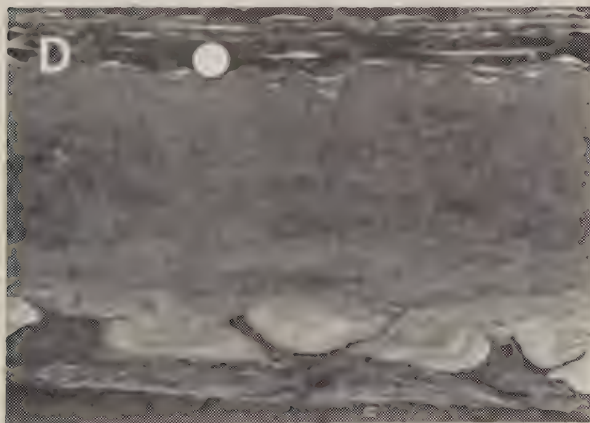
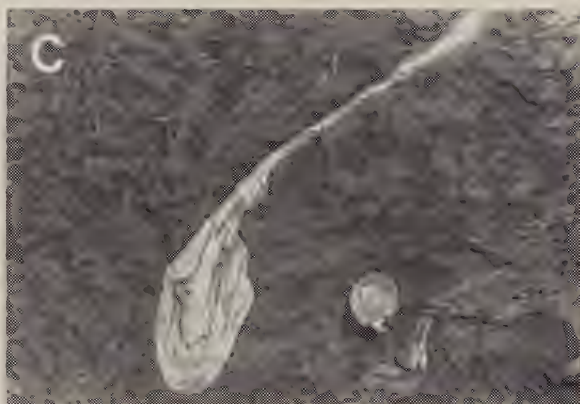
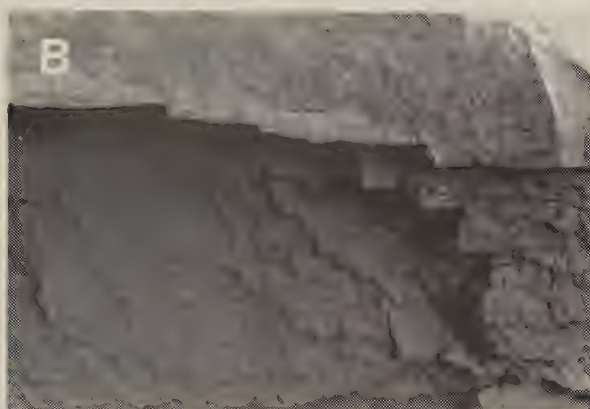


Fig. 80. Deformed and reworked silt and clay sediments of unit 1a:8. A. A slump scar with folded and thrust layers (zone A) grading upwards into homogeneous silt (zone B). B. Homogeneous silt of zone B. Note the horizontal splitting. C. A pouch of laminated silt which has sunk into the homogeneous part. D. A complex unit, comprising zone A (deformed and partly dissolved sediments) zone B (homogeneous silt) zone C (zone of load casts) and zone D (cross-laminated silt). E. A bed-set of sharp-crowned symmetrical ripples, dissolved by liquefaction in the lower part.

result of loading from rapidly deposited sediments or by phenomena as earthquakes and storm waves (Rupke 1978).

The large variations in the thickness of clay laminae rather indicate irregular time intervals between the different turbidity underflows.

The sedimentation of the middle and upper part of the silt and clay facies of unit 1a:8 was affected by fall-out from suspension, wave action, slumping and gravity flow.

The clay laminae are the result of fall-out from suspension during very calm conditions. The ripple sets of symmetrical ripples were deposited during wave action at a more or less continuous sediment supply.

Slump scars and thick units of deformed and dissolved strata with different degrees of destruction of the internal coherence show slumping activity in the basin. All degrees of reworking are present in the sediments, from slightly deformed strata close to the slump scar, to debris flows with different degrees of internal shearing and to turbidity flows.

The thick strata of silt and clay (10.5 m) cannot be explained only by intraformational redeposition by slumping and wave action. A contribution of material into the basin must also have occurred. The absence of sand size particles indicates that the particles were suspended in low density turbidity currents, probably as a more continuous supply of suspended sediments, and as steady turbidity currents of low density (Friedman and Sanders 1978, p. 512).

Structures as graded bedding, asymmetrical ripples, parallel-lamination, which normally are connected to turbidity currents, are quite rare. This may depend on several factors. Reworking by wave action and slumping may have destroyed the turbidities. The reworking processes may, however, produce new turbidities.

Another factor, which prevents the formation of turbidities, is the character of the inflowing turbid water. Very small density differences between the inflowing turbidity current and the surrounding water masses may result in overflows or interflows (Gustavsson 1975), or in a rapid mixing between inflowing streams and surrounding waters, producing unstratified water (Sturm 1979).

A rough estimation of the sedimentation rate in the basin can be made even if a cyclic deposition is not evident. The maximum thickness of clay laminae in the silty part of the sequence is 1.5 cm. If the wave activity is considered an annual occurrence, the value can be used as a very approximate thickness of the winter accumulation. When transferred to the lower clayey part, an annual sedimentation of 1 to 1.5 cm gives a time span of 100-150 years for the sedimentation of the 1.5 m thick sequence. A comparison to clay laminae of distinctly varved sediments in other stratigraphic positions in the area gives similar winter accumulation rates. The clay laminae from varved sediments in other parts of Sweden often fall in the range between 0.5 and 1.5 cm.

The approximate sedimentation rate in the silty part is based on the average thickness of silt beds with primary sedimentary structures between clay laminae. The thickness ranges between 10 and 15 cm. This gives a sedimentation time of 60 to 90 years for the 9 m thick silty part.

Log 2

This section (log 2, Fig. 81) is situated only about 300 m to the south of section 1. Nevertheless there are significant differences in sedimentary characteristics, important for the general environmental interpretation.

Sedimentary characteristics

The transition from the Ålabodarna member is gradual without erosional contacts, and the stratified diamicton grades into silt and clay facies of the Glumslöv member.

The silt and clay is varved with distinct boundaries at the silt/clay transitions and the clay/silt transitions. 38 varves were recognized. The clay laminae range from 0.2 to 1.0 cm in thickness. The same division into a lower brown-gray part and an upper dark-red part as in section 1 (1a:3) was observed. The main part of the silt layers are parallel-laminated, the individual laminae being approximately between 0.05 and 0.1 cm thick.

The varve thickness increases upwards from generally less than 2 cm in the lower part to a maximum of 5 cm in the upper part. The uppermost nine silt layers are cross-laminated. A combination of A-type ripples, grading into B-type ripples and draped by parallel-laminated silt, is most frequent. Three of the silt layers consist of multiple ripple sets. A 40 cm thick bed of homogeneous and convoluted silt lies between varve 29 and varve 30.

Mean grain sizes in the silt layers range between 5.5 ϕ and 6.9 ϕ . The samples contain several thin laminae and represent bulk samples in the silt layers.

The silt and clay facies gradually changes into sand facies or rather into coarse silt and sand facies. Mean grain sizes generally lie between 3.5 ϕ and 5 ϕ with a slight increase of average grain size composition in the middle part of the 5.5 m thick sand sequence with a maximum mean grain size of 2.95 ϕ . Only one clay lamina could be continuously traced in the cutting. However, some of the cross-laminated sets or cosets contain a high amount of grains or flakes of intraformational reworked clay (Fig. 82A).

Ripple cross-lamination is the most abundant structure, and the ripples in the coarse silt and very fine sand show a large variety both in ripple forms and ripple sizes (Fig. 82B). The silty lower part is dominated by symmetrical and asymmetrical ripples, generally with very low values of vertical form index (4.5 to 8.5). The symmetrical ripples are generally sharp-crested with rounded troughs (Fig. 82C). The pointed crests lie either in phase or displaced, often with a gradual change in angle of climb. Wavy erosive contacts between different ripple sets have not been recognized and seem to be absent. The asymmetrical ripples are either of ordinary A-type or large-scale ripples with superimposed micro-cross-lamination, similar to those described by Stanley (1974). The ripple foresets are generally directed towards the north-west, but the sequence is intersected by a few thin sets of asymmetrical ripples, directed reverse to the general trend.

The grain size parameters in the silty sand and sandy silt ripples show standard deviation (1) values between 0.45 and 1.68. There is a distinction between A-type ripples with a better sorting and B-type ripples

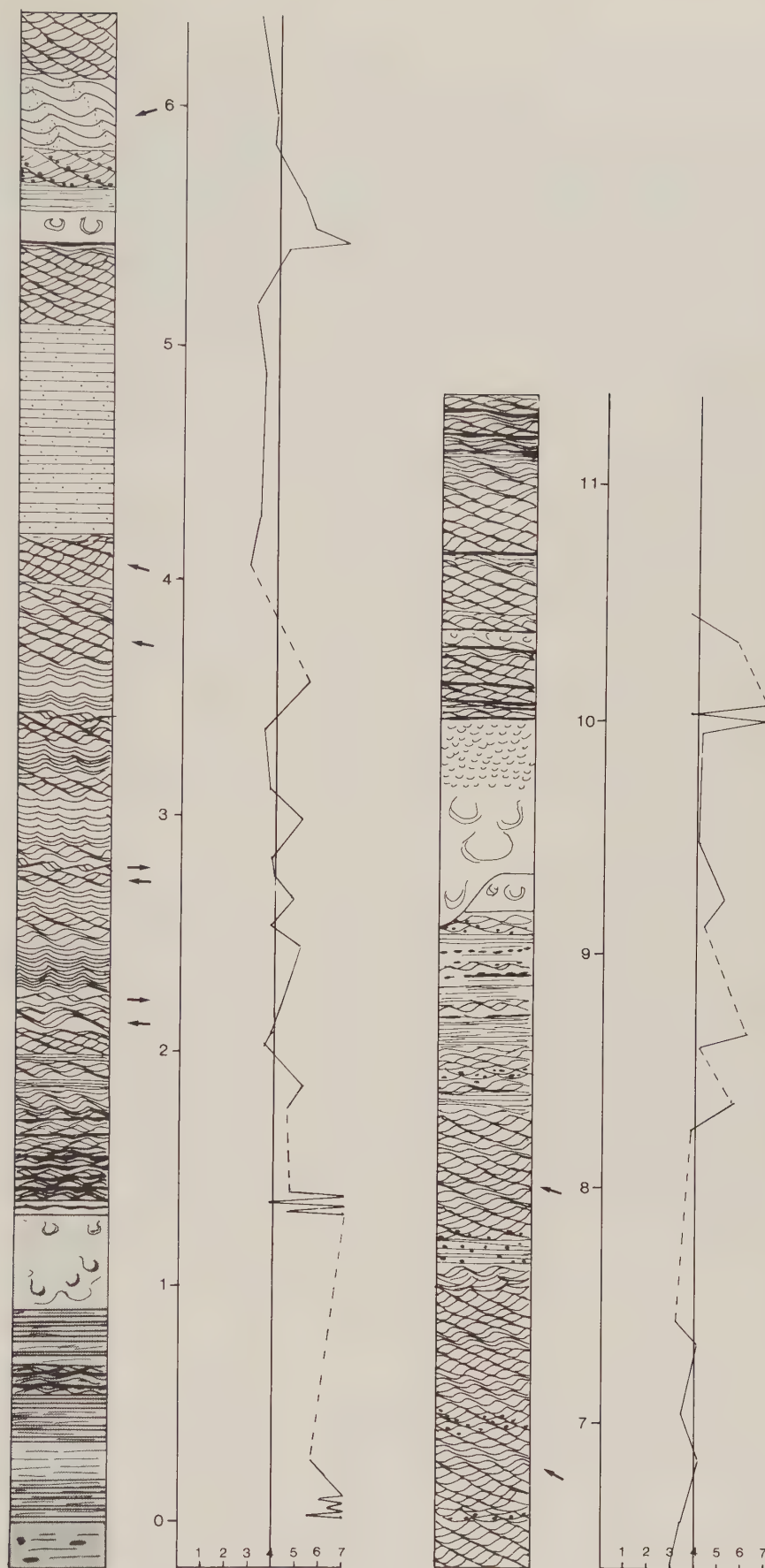


Fig. 81. Log 2, comprising units 8, 10 and 11 at 1:475-485.

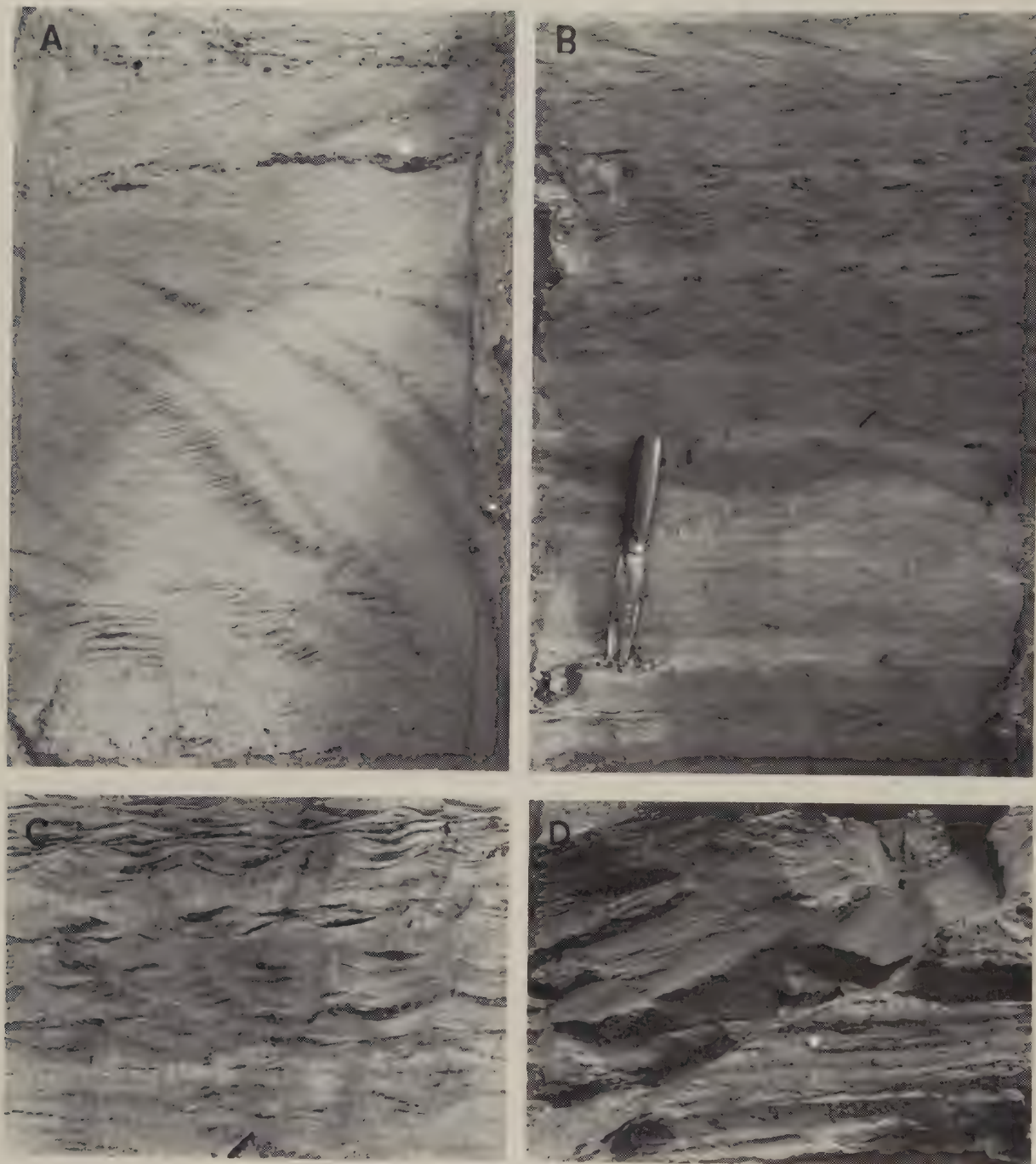


Fig. 82. Sedimentary structures from log 2. A. Ripple cross-laminated sets with widely spaced zones with clay intraclasts. B. Current ripples and oscillation ripples. The arrows indicate ripple sets directed reverse to the general trend. C. A close-up of the sharp-crested symmetrical ripples. D. Channel scour. The infilling is silty sand with convolute and dish structures.

and sharp-crested symmetrical ripples with a moderate or poor sorting. Skewness values are less varying and range between 0.26 and 0.45.

The middle and upper part is dominated by A-type ripple lamination in tabular sets, generally with a very low inclination of ripple climb. Planar lamination is also present in the middle part. The sand contains spaced parallel zones with intraformational clay clasts. The vertical distance between the clay clast zones decreases upwards, and strongly eroded clay and silt layers in primary position turn up. The sediments

grade into alternating bed facies.

Between 6.95 m and 7.9 m there are two thick beds with convolute lamination. A 1 m deep channel scour is infilled with silty sand with convolute structures in the lower part and dish structures in the upper part (Fig. 82D). The layer is not restricted to the channel but can also be traced outside the channel in a 30 cm thick bed, lying conformly on the substratum.

The alternating bed facies continues on top of the convoluted beds. Distinct clay laminae with sharp upper and lower contacts sandwich cross-laminated sets of

very fine sand. The thickness of the sand layers ranges between 2 cm and 45 cm.

At 10 m the sediments become strongly involved in glaucitization, and shearing and thrusting made further logging impossible. The lower part of the strata, involved in the subhorizontal thrusting and folding, consists of silt and clay facies, and it is likely that the alternating bed facies is overlaid by laminated silt and clay.

Interpretation

The varves in the lower part of the section show a cyclic repetition of different sedimentary processes. The clay is the result of a slow settling of particles in a basin of still-water. The laminated silt is the distal part of a low density turbidity current. The addition of a C-division of the Bouma sequence in the upper varves is probably the result of increasing proximity to the source of the turbidity currents. The varves are similar to those described as Group III varves by Ashley (1965) with silt layers thicker than clay layers. This type of varves was formed in the most proximal position, related to the inflowing rivers.

The structures in the sand sequence indicate both unidirectional currents and oscillatory movements. The sharp-crested symmetrical ripples were formed by applied shearing stresses in two opposite directions as a back-and-forth oscillating movement. The ripple laminae in phase or steeply climbing ripple sets were formed during a more or less continuous sediment supply from suspension (Walker 1963, Mc Kee 1965, Jopling and Walker 1968, Allen 1973). This process can be denoted to different environments. Symmetrical ripples are frequent in sediments, influenced by oscillatory wave motion. However, the symmetry does not necessarily imply that the ripples were formed by wave action (Jopling 1961, Mc Kee 1965, Boersma 1967, Aario 1972 a, b, Sundborg 1956, Mc Kee 1938), and symmetrical ripples have also been found in fluvial environments. The symmetry does not necessarily imply wavy action.

The formation of symmetrical ripples, created by the orbital movement from the passage of wind-generated waves, was described by Bangold (1946). Such ripples, when formed in silty sediments, can be found in the low velocity area at lake or sea floors in the outer part of the shoaling-wave zone, between wave base and transformation base (Friedman and Sanders 1978 p 575). Sharp-crested symmetrical ripples in the fluvial environment are connected to the zone of zero velocity which arises in connection with boundary layer separation at the foreset top of a small delta or a bar (Jopling 1961, 1965, Boersma 1967).

Aario (1972 a,b) doubted the possibility to separate the wave-generated symmetrical ripples from the symmetrical ripples in the zero-zone of the distal part of the delta (Aario 1972a, p 157). The fluvial character could only be assigned if an association of other bedforms from the delta system were recognized.

The sedimentary conditions in the two environments are, however, different, and important dissimilarities could be recognized. The first important difference is the access of grain sizes at the bed. Ripple sets in phase or with a high angle of climb indicate a more or less continuous sediment supply from above as a fall-out from suspension.

The velocity in the orbital water movement of an oscillatory wave decreases downwards in the water column, and the motion keeps the fines in suspension. The grain size distribution of particles, reaching the bed, is thus impoverished of fines. The oscillatory-elliptical motion close to the bed is a continuous movement, which prevents the sedimentation of fines and the sediments become coarse-skewed. Only when wave action ceases the fines settle down and drape the ripples with a clayey lamina.

The sediment supply to the symmetrical ripples in the fluvial environment comes from suspended particles in the jet, leaving the top of the delta and entering the zone of mixing. The proximity to the delta gives a very high rate of sediment supply, and the sediment particles are continuously moving through the locus line of zero velocity with decreasing maximum grain sizes at increasing distance from the delta foreset. When the zone of zero velocity attaches the bed, all the suspended particles, also the finest, are trapped in the zone and submitted to sedimentation. The oscillating water agitation is a slow back-and-forth motion parallel to the bottom, which involves a momentum of stand-still at every turnpoint, which permits a fall-out of the smallest grains. The grain size composition of symmetrical ripples in the bottomset of a delta is characterized by positive skewness.

Another differentiating aspect is the continuity and the steadiness of the processes during the aggradation of the sediments. The shear stress at the bed under a wind-generated wave is dependent on the size of the wave, which in turn is governed by the speed of the wind, the duration of the wind and the length of the fetch. These components are extremely varying, and during the short summer season when the glacial lake surface is exposed to wind action, one could expect numerous variations between stormy, windy and calm weather, resulting in alternating erosion, ripple aggradation and fall-out from suspension. Shearing stresses thus vary from values high enough to erode the lake flow to zero when no motion occurs, and only a fall-out from suspension affects the floor. Symmetrical ripple sets are thus frequently interrupted by wavy erosive surfaces and by draped lamination.

The changes in sedimentary conditions at the delta bottom set are mainly connected to changes in the velocity of the jet. The position of the attachment point may vary laterally, and a certain point at the bottom may be affected by reverse circulation, zone of zero velocity and zone of coflow. The energy variations in the distal delta environment are most probably less varying than lake bottom areas, affected by wind-generated waves. The ripple cosets in the sand sequence are interpreted as bottom set deposits because a) the sediments are fineskewed, b) the sediments lack erosive contacts and represent a continuous aggradation and c) the sediments show asymmetrical ripples in two opposite directions in combination with symmetrical sharp-crested ripples, reflecting the flow conditions at the attachment point.

The planar laminated sand in the middle and upper part of the sand sequence indicates upper flow regime conditions. Zones of clay intraclasts, grading into eroded layers in primary position, show that the water altered between high flow velocities and complete stagnation. The current structures

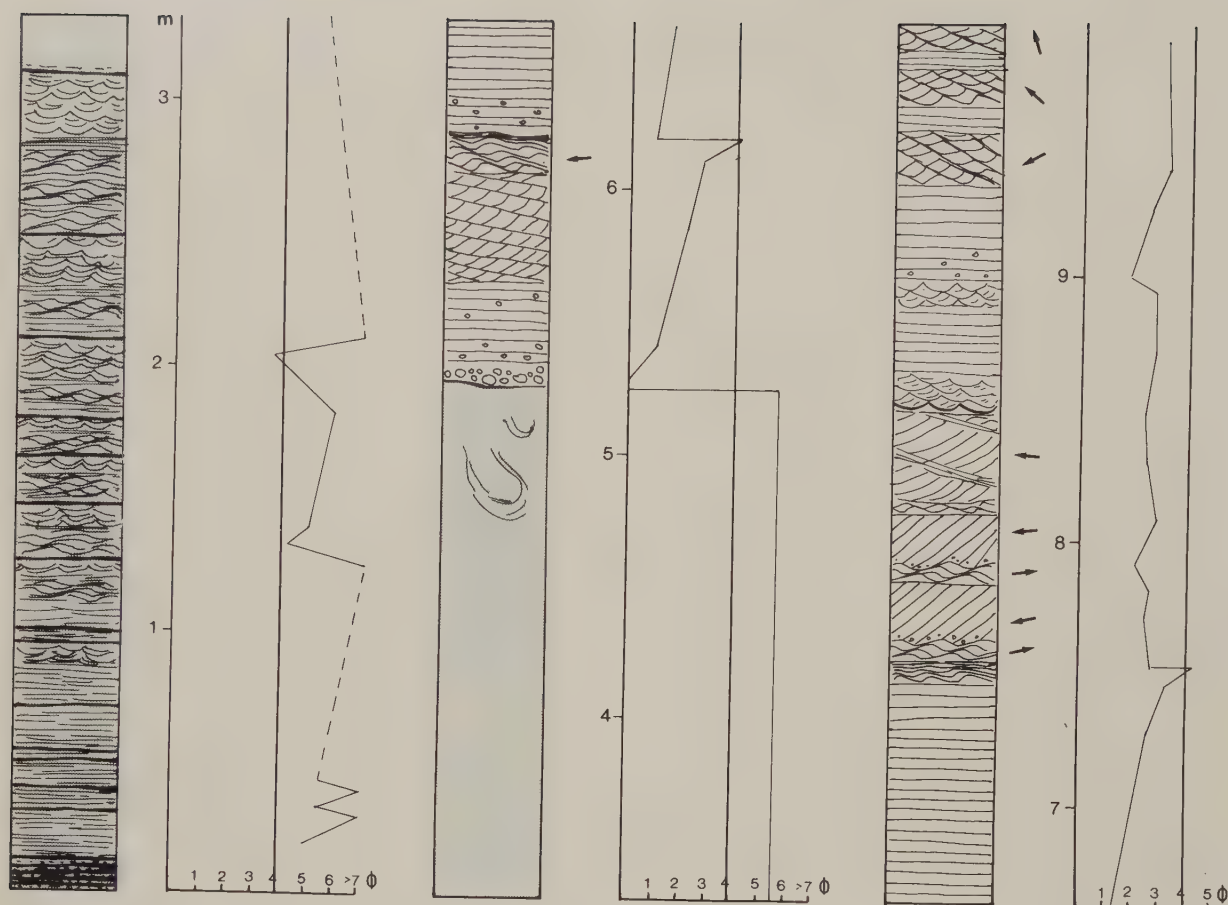


Fig. 83. Log 3, comprising units 8.9 and 10 (Fig. 62) at 4:230.

were formed from the stream, which discharged into the lake and followed the lake floor as an underflow current.

The channel scour and the thick units of convoluted bedding can be connected to gravity flows on the delta slope. The high density flow originated from sediments already deposited, which become liquefied either from the shock of an overriding current or spontaneously from unstable conditions on an oversteepened slope. The dish and pillar structures in the silty sand were formed by plastic deformation during the dewatering and settling of grains simultaneously to the deposition. The sequence shows a gradual transition from a glacial lake sedimentation to a sedimentation in a pro-delta environment. A decreasing influence of high velocity flows in the uppermost part and a likely upward transition into silt and clay facies indicate a new phase of glacial lake sedimentation. The place never emerged over the lake surface.

Log 3

The investigated sequence (Fig. 83) is situated at the inner wall of the landslide scar at 4:230. The sequence comprises units 4b:3, 4b:4 and 4b:5.

Description

The silt and clay facies (unit 4b:3) in the lower part of the sequence is a varved sediment with distinct clay laminae, alternated by silt laminae or layers. 18 varves were exposed in the section, but the varved sediments continue below the excavation.

The clay laminae range in thickness between 0.7 and 1.5 cm. The upper and lower boundaries of the clay laminae are distinct. The internal structure is similar to the clay laminae of unit 1a:4, described in log. 1.

The silt layers increase in thickness

from about equal to the clay laminae in the lower part to a maximum of 40 cm in the upper part. The silt layers in the lower varves are parallel-laminated and consist of thick graded laminae, generally about 1 cm thick, and micro-laminated silt (Fig. 84). Fine sand streaks occur between the silt laminae. The silt layers in the upper varves consist of parallel-laminated silt and cross-laminated silt. The cross-laminated parts have symmetrical ripple sets with draped parallel-lamination. The ripple sets are generally in phase in the lower part and grade upwards into form-discordant sets with wavy erosion between the individual sets.

The silt and clay facies ends up with a 2.8 m thick layer of massive and convoluted silt (B-C -type).

The sequence starts with a widespread pebble-strewn surface, which may increase to a 25 to 30 cm thick layer of poorly sorted sandy pebble (unit 4b:4). The pebble is massive or with a faint horizontal bedding. Small lenses of sorted sand occur. Clay and silt intraclasts make up 10 to 50 % of the pebble sized particles. Several intraclasts have a partly sorted composition. The lower

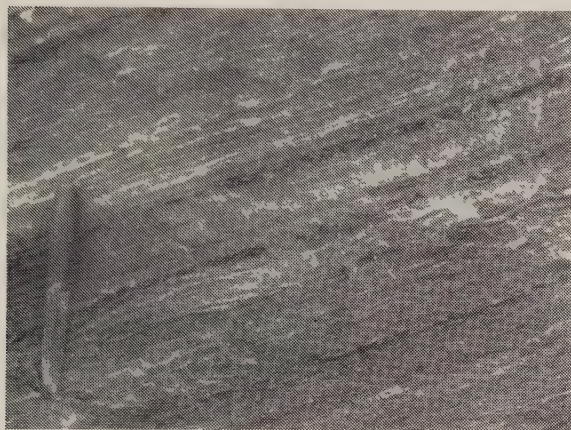


Fig. 84. Silt and clay facies. Clay laminae with distinct upper and lower boundaries. Graded bedding, micro-lamination and fine sand streaks in the silty sets.

boundary of the pebble layer is slightly scoured. Some minor flutes or pools, rarely exceeding a depth of 5 cm, were observed on the lower silty clay surface. Large-scale trough or channel-formed scour marks were absent.

The pebble grades into coarse sand with planar lamination and tabular sets of A-type cross-lamination. The next bed is a very fine sand with B-type cross-lamination, which grades into coarse silt with sinusoidal ripple lamination and draped lamination. These beds, from the scoured surface to the top of the draped lamination, make up a 90 to 100 cm thick upward sequence.

The crests of the ripple-formed surface have been planed off and overlayers by planar bedded sand that makes up the major part of another fining-upward sequence, which is concluded by 2.5 cm of cross-laminated fine sand and draped laminae of silty clay. The sequence is 135 cm thick.

The middle part of the sand is dominated by cross-bedding. The first bed on top of the fining-upward sequence is a 3 to 5 cm thick coset of A-type ripple cross-lamination. The top of the coset is eroded or dissolved in a thin bed of massive sand. The massive sand is interwoven with a 25 cm thick cross-bedded unit. The upper surface of the unit is plane, and diffuse planar laminated sand touches the surface with a low-angle dip of 5°. The inclination of the laminae increases and grades into a foreset slope with a maximum inclination of 25-30°. The foreset laminae about the floor tangentially. The foreset laminae are graded, and coarse sand grains are concentrated at the toe.

The relationship between different bedforms in the next interwoven tabular coset is easily interpreted from Fig. 85 where a dark mineral lamina is draped on the foreset slope of the cross-bedded set and on the ripples at the toe of the foreset. The A-type ripples are directed opposed to the major foreset slope. The laminae in the foreset bed are planar and have angular contacts to the bottom set. The foreset bed is 15 cm thick. The foreset laminae dip 22°.

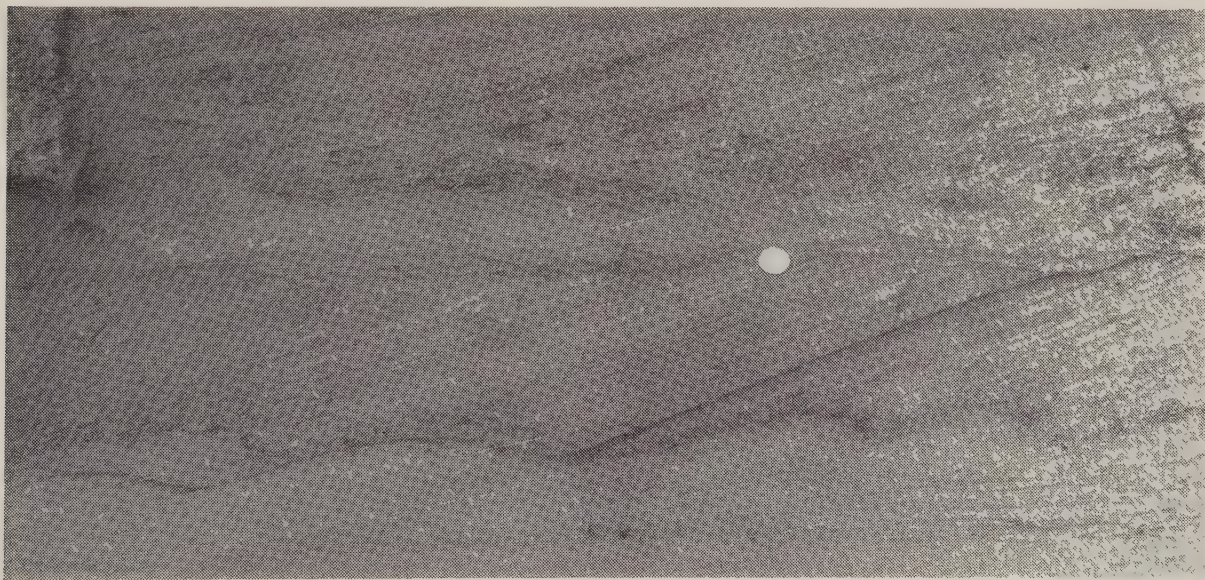


Fig. 85. A tabular set of cross-bedded sand with regressive ripples in front of the foreset slope. The dark laminae shows contemporaneous bedforms.

Next unit is a 40 cm thick coset of trough cross-bedding. Individual sets have an average maximum thickness of 15 cm. The upper surface of the individual sets is erosive and has a dip of 5 to 7° in upstream direction. The foreset laminae have a maximum inclination of 15° (downstream direction), and the contact to the bottom set is tangential. The bottom set consists of a crudely ripple-laminated or massive fine sand. Very coarse sand particles are scattered in the foreset laminae with a slightly increased concentration at the toe.

The total thickness of the cross-bedded cosets is approximately 1 m. The upper part of the sand sequence consists of an alternation between planar bedded sand and cross-laminated sand of A-type ripples in tabular climbing sets and in trough-shaped sets. The grain size composition is generally fine to very fine sand with increasing grain size composition in some of the planar laminated sets.

Interpretation

The varved sediments in the lower part of the sequence show a cyclic alternation of sedimentary process. The clay laminae were formed by fall-out from suspension in a calm water environment. The parallel-laminated 'summer'-layer in a varve couplet is generally considered deposited from turbidity currents (e.g. De Geer 1912, 1940, Kuenan 1951). Parallel-laminated silt is associated with the Bouma sequence (Bouma 1962) where it constitutes the D-division and often forms the only primary sedimentary structure in combination with the E-division (Bouma 1972, 1973).

The origin of these parallel-laminated muds have been discussed, and several interpretations are suggested, e.g. (1) Pulsatory velocity fluctuations cause alternation between bedload and suspended load sedimentation (Piper 1978). (2) A silt lamina is deposited when bottom shear stresses reach the limit for sedimentation. This implied an increase in clay concentration in the suspension and results in a clay flocculation and sedimentation (Stow 1977, Stow and Bowen 1978). (3) The "burst-and-sweep model" for the viscous sublayer of turbulent flows, as described by Jackson (1976) and Bridge (1978), has also been applied to laminated silts by Hesse and Chrough (1980). Shear sorting results in winnowing of clay fractions and sedimentation of silt during "burst-and-sweep" events when a turbulent vortex reaches the sublayer, and silt and clay sedimentation during intervals when low speed streaks prevail (Hesse and Chrough 1980). Whichever of the explanations is chosen, it demands grain movement and shear stress along the bottom.

The varve thickness increases upwards. Ashley (1975), in his investigation of glacial lake sediments, shows a direct relationship between varve type and varve thickness and the proximity to inflowing rivers with increasing silt layer thickness at decreasing distance.

The sharp-crested symmetrical ripples in the upper part of the varved sequence were formed by oscillatory flow movements. A prevailing characteristic is that the influence of wave action is restricted to the upper part of the silt layer, where the oscillatory ripples alternate with wavy, erosion, draped lamination and parallel lamination. That means that the summer season started with

sedimentation from turbidity currents. The thick beds of symmetrical ripple sets, sometimes with ripple in-phase formed sets, indicate a wave action during more or less continuous sediment supply. The upward increase of form-discordant sets indicates a decrease in sediment supply at the end of the summer season before the lake surface was covered by winter ice and the sedimentation changed into a slow settling of clay particles and a clay lamina was formed.

The reason why the wave action did not affect the silt in the lower varves is probably due to a position of the glacial-lake floor below the wave base. The increase in silt layer thickness and the appearance of oscillation ripples in the upper varves can be connected to a shore advance and increasing proximity to the inflowing river mouth, consequently leading to a more proximal varve facies and a transition to a bottom position above the wave base, which was the result of a lowering of the glacial lake level.

The massive silt with ball-and-pillows of load cast type lies with non-erosive contact to the uppermost clay laminae. The sediment was brought into the area as a high density gravity flow with a most probable source in a slump of silty glacial lake sediments, connected to the lowering of the glacial lake level.

The high amount of intraclasts in the pebble-strewn surface on the top of the silt shows that the deposition was preceded by erosion. The pebble-strewn surface or thin pebble bed is laterally extensive, and there is no evidence of large-scale scour marks or channels. The low relief of the extensive scoured surface does not, accordingly, indicate deposition in deep channels. The lateral spreading was instead facilitated by a fluvial pattern with poorly defined shallow channels, which migrated laterally over the flat area. The coarse bedload was transported in the central part of the channels where competency was greatest, and the bedload was deposited either as a lag in scour pools (D.G. Smith 1973) or as the result of reduction of competency where a channel widened, and a small submerged central bar was deposited (Hein and Walker 1977). The small sand lenses in the coarse-grained basal part of the sand sequence indicate that the deposition was not a virtually simultaneous event but rather a combination of several depositional processes in restricted patches or pools and separated in time and space. The cross-laminated sand on top of the pebble and planar laminated pebbly sand indicates a reduced flow intensity. The ripple bedform was produced in the lower flow regime (Simons and Richardson 1961, Simons et al. 1965). The B-type ripples and the draped lamination show a further slowing down to a finally standing water. The planar bedded fine sand in the second fining — upward sequence indicates upper flow regime stream (Simons and Richardson 1961, Simons et al. 1965). A thin set of cross-laminated fine sand (lower flow regime) was deposited before a final stagnation of the flow and deposition of silty clay.

The first cross-bedded unit is a washed out dune, a transition form from lower flow regime dune bed to upper flow regime plane bed. The avalanching of material down a foreset slope of a dune is replaced by an almost continuous sheet flow of sediment. The steep avalanche face gradually decreases in height in flow direction and is replaced

by the faintly laminated, slightly dipping upper part of the foreset slope. The structures show a transition into a very high depth-ratio (Jopling 1965). This effect was experimentally obtained by Simons, Richardson and Nordin (1965) by an increased flow velocity over a dune bedform in a flume.

The second cross-bedded unit is a tabular set which generally is connected to linguoid or transverse bars (Collinson 1970, Smith 1970, 1971, Cant and Walker 1978). The foreset has an angular base which is formed beneath a reduced separation eddy (Collinson 1970). Angular toesets are favoured by a low depth ratio of the stream over the bar, related to the depth at the toe, or low flow velocities (Jopling 1965). The graded foresets show that avalanching was the dominating process in the lateral migration of the bar (Allen 1965).

The dark laminae which drape the foreset slope and the regressive ripples in the bottom set layer must, however, have been derived from suspended material. An alternation between avalanching processes and fall-out from suspension was probably responsible for the distinct lamination of the foreset bed. Such an alternation may be the result of pulses in sediment transport, related to the moving of bedforms along the stream bed (Jopling 1966), or it may be the result of velocity pulsation in the stream or in the viscous sublayer of turbulent flows (Jackson 1976, Bridge 1978). The small height of the foreset slope, the angular contact to the bottom set and the tabular form of the set is interpreted as a straight-crested bar, developed in a shallow channel as a linguoid or transverse bar (Smith 1970, Southard 1975, Miall 1977).

The trough-shaped upper cross-bedded sets are the result of three-dimensional dune bedforms, which are produced at increasing flow velocities (related to the sandwave) in the upper part of the lower flow regime (Southard 1975). Dune bedforms are frequently found on straight-crested bars and are formed as dune accretion on the bar surface (Collinson 1970, Miall 1977).

The alternation of cross-lamination and planar lamination in the upper part shows repeated transitions between lower flow regime and upper flow regime. The palaeocurrent direction, based on orientation of ripple foresets, shows several changes in flow direction. It is possible that the surface emerged between the different overflows even if the 140 cm thick sequence shows no real signs of interruption in the sedimentation.

A summary of the environmental interpretation is: the silty surface was invaded by fluvial currents. A shallow channel pattern was developed. Two fining-upward sequences indicate an intermittent flow. The cross-bedded sequence was deposited in a shallow channel by the migration of dunes and bars. The water depth on the top of the bar was decreasing, and planar laminated fine sand was deposited. It is probable that the bar surface emerged during falling stage and developed into a sand flat. The bar surface again submerged during new flood stages but probably only with shallow sheet flows, which deposited planar laminated sand and ripple cross-laminated sand at decreasing flow intensity.

Log 4

The log (Fig. 86) was recorded in the most southern section on the mainland. The strata are dislocated, and the originally horizontal beds have obtained a general dip of 20° to 25° towards the south (S22°E to S6°E). A lateral transfer was necessary in order to reach all the strata in the section.

Description

Four main facies states were recognized (chapter 5.1.16): a lower silt and clay facies (0–9 m), a sand facies (9–19.5 m), alternating bed facies (19.5–23.8 m) and an upper silt and clay facies (23.8 to 28 m).

The lower silt and clay facies consists of distinct 0.5 to 1.8 cm thick clay laminae, which lie as continuous parallel beds, sandwiching the silt layers and thick units of deformed or convoluted beds. The clay laminae have sharp upper boundaries. The lower boundaries are either sharp or gradational with a graded transition zone less than 1 mm thick. The clay content in bulk samples from six different laminae exceeds 80 %. The clay laminae have a lower part with lighter shade than the upper part, and the silt content seems to be slightly higher in the lower part.

Clay laminae are also folded and wrinkled in some of the deformed units.

Undisturbed primary sedimentary structures in silt layers occur between 5.1 m and 5.5 m and consist of cross-lamination and parallel lamination, either as draped laminae on top of ripple sets or as horizontal laminae.

The thick silty beds, which range in thickness from 40 cm to 150 cm, are composed of (from bottom to top) B-D zones, B-C-B-C-B-C-B-C-B-D zones, B-D zones, B-C-B-D zones, A-D zones, A-B-C zones and B-D zones. The uppermost unit consists of sandy silt, which is convoluted in the lower part and has dish structures in the upper part (Figs 87A and B).

Mean grain sizes from samples in the sand sequence fall within fine sand and very fine sand fractions and range from 2.18 to 3.75 ϕ .

The predominating structure in the 10 m thick sand sequence is tabular sets of cross-bedded sand either in cosets or interlayered by ripple cross-laminated or planar laminated sand. The bounding surfaces of the cross-bedded sets are generally parallel and planar, and some sets were traced for several metres. A few sets were wedge-formed with a slight increase in foreset bed thickness in downstream direction. One of the sets started as a cross-laminated set, which grew into a large-scale cross-bedded set.

The individual foreset beds range in thickness between 30 cm and 150 cm. Thicknesses between 50 and 100 cm are most frequent. Both angular and tangential contacts between foreset laminae and the floor occur. Bottom sets with regressive ripples were observed in two cases.

The approximately 1 m thick bed at 19 m consists of very fine sand and coarse silt and has convoluted lamination in the lower part and dish structures in the upper part. This bed marks the transition from sand facies to alternating bed facies.

The alternating bed facies consists of between 30 and 35 fining-upward sequences, with a maximum thickness of 50 cm and an average thickness of 10 to 15 cm. Planar-

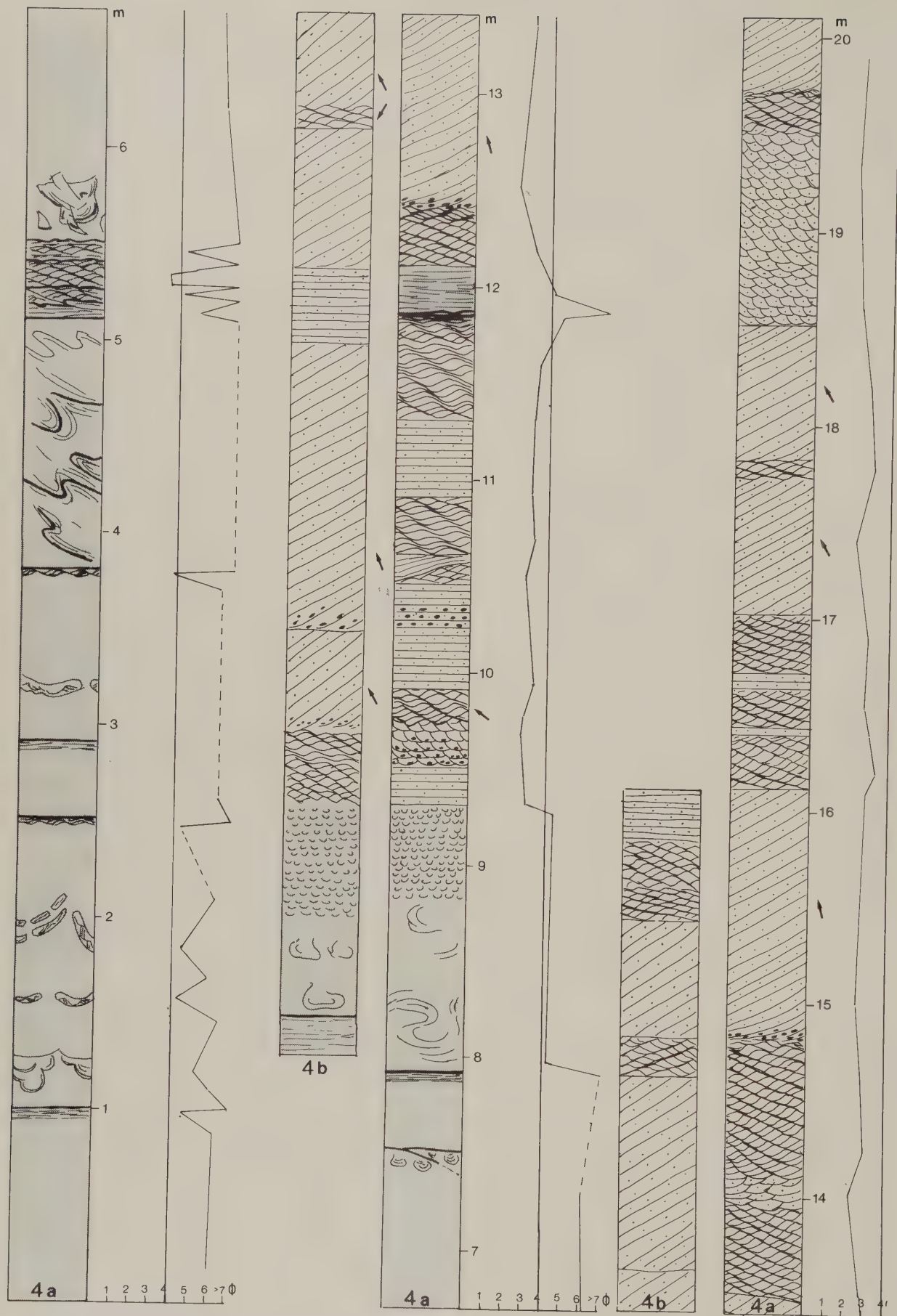
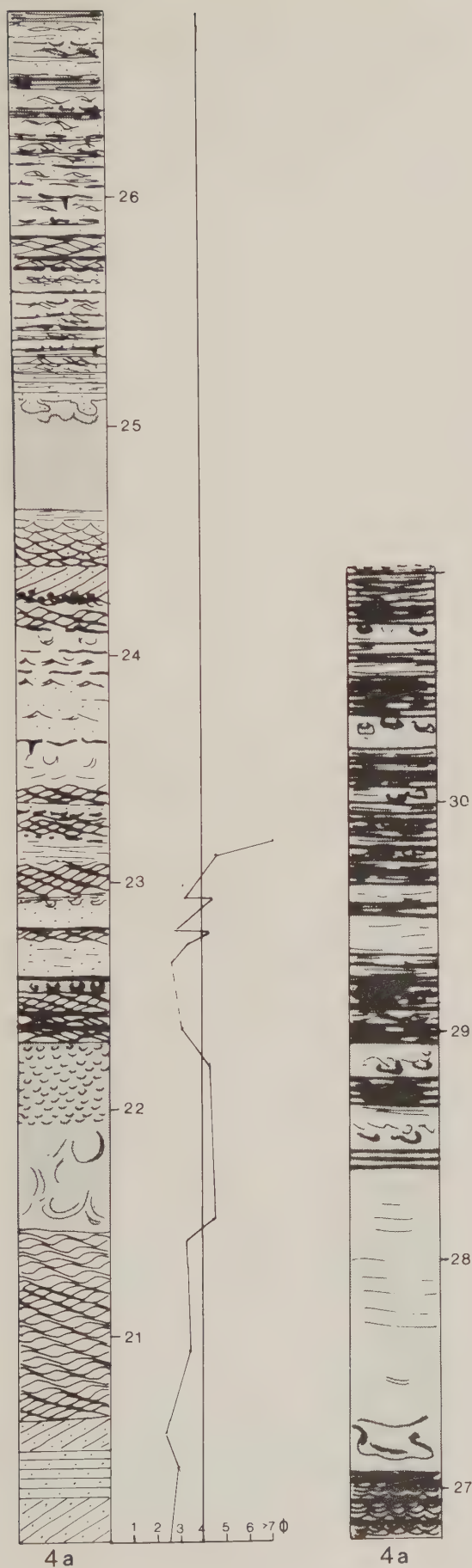


Fig. 86. Log 4, comprising units 8, 10, 11 and 12 (Fig. 62) at 16:770-850. Log 4b is recorded parallel to log 4a from 8 to 16 m. The two measures are 10 m apart.



laminated sand, A-type ripple lamination, B-type ripple lamination and finally draped lamination of coarse silt, grading into a very thin clay lamina, were recorded in some of the fining-upward sequences. The clay laminae, which generally are less than 0.1 cm thick, are often broken to pieces and rolled up in the edges. Primary sedimentary structures in the silt and sand beneath the clay laminae are frequently partly destroyed or dissolved. Small cracks or fissures, generally less than 10 cm deep, are developed from the clay surfaces. The alternating bed facies on top of the sand facies was also exposed in a minor section in a ravine approximately 100 m to the north of the investigated section. These strata contain some well-developed ice-wedge casts (Fig. 88).

The sand ceases at 21.2 m. and the strata continue as an upper silt and clay facies. Parallel-lamination, asymmetrical ripples and symmetrical ripples with sharp crests are present in the silt layers. Clay laminae are draped on wavy ripple surfaces or lie as horizontal layers. The thickness of the clay laminae ranges between 0.2 and 1.5 cm. The lower part of the silt and clay facies contains two beds of deformed and convoluted strata. The 150 cm thick unit between 21.4 m and 22.9 m comprises A-B-C zones, and the upper 13 cm thick unit at 23 m consists of C-D zones.

The silt layers in the topmost part decrease in thickness and are sometimes developed as extremely thin silt streaks, sandwiched by clay laminae. The symmetrical ripples

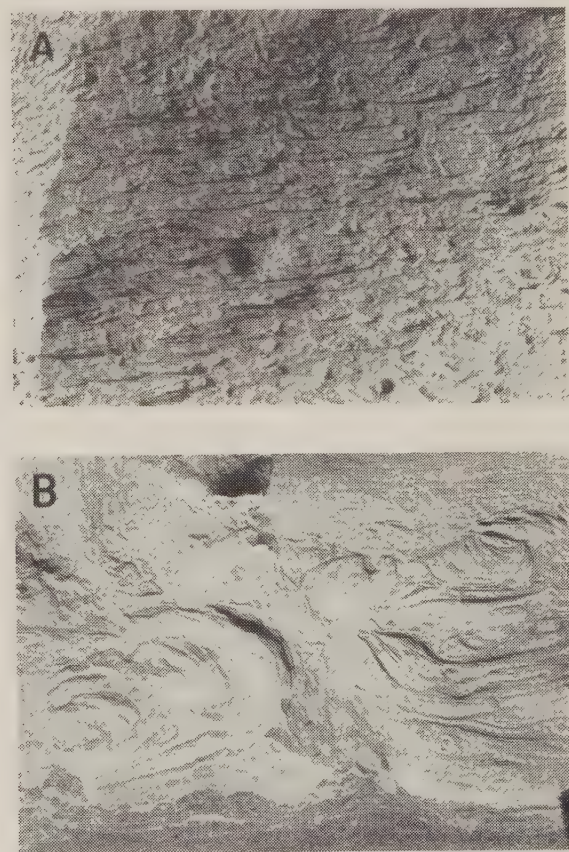


Fig. 87. Convolute bedding (B) and dish structures (Fig. A) in silty sand at 8 to 9 m level in the log.

disappear upwards, and also asymmetrical ripples become rare and often appear as lenticular beds, draped by parallel-laminated silt or clay laminae.

Interpretation

The silt and clay facies in the lower part of the sequence consists of clay laminae, formed by fall-out from suspension, and silt layers, deposited from sediment gravity flows. The various structural types of the silt layers represent a series of mass-gravity transport with differences in water content.

The silt layers at 5.10 to 5.50 m are composed of cross-laminated silt with an upward grading into draped and parallel lamination. These layers, the C and D parts of the Bouma sequence (Bouma 1962), represent a deposition from the bedload of low concentration turbidity currents.

The silty layers at 0 to 3.8 m and 6 to 7.7 m are composed of a thick lower massive part and an upper part with ripple cross-lamination and parallel lamination. In a few cases, the unit between two clay laminae consists of multiple layers, each one representing a sedimentation event. The massive lower part represents a rapid deposition from a high density turbidity current and is followed by ripple formation in the lower part of the lower flow regime and finally draped by silt laminae in the waning stage of the density current. The multiple layers at 1 to 2.5 indicate a repeated pattern of non-uniform high density currents.

The sandy silt in the upper part of the unit has settled out from a liquefied flow. The dish structures were formed during the upward escape of water (Lowe and Lopiccolo 1974, Lowe 1975). Finally, the sediments between 3.8 and 5.1 m, and 5.5 and 6.0 m consist of slumped sediments. Primary sedimentary structures are preserved from the original sediment, and the sedimentary structures are only partly destroyed and dissolved.

It is widely held that the source of high density flows is intrabasin slides and slumps (Morgenstern 1967). Slumping is frequent in areas of rapid deposition (Moore 1961) and is a common process in the spreading of delta slope sediments into a basin. The low density currents may be developed from the tail of a high density turbidity current but may also come from a muddy stream, entering the basin.

The sand sequence on top of the silt and clay sediments is produced by bedload deposition in unidirectional flow, and the fluvial character of the sediments is obvious. The transition from silty sediments to sand with a fluvial origin is rapid and without indication of a prodelta foreset slope of Gilbert delta type (Gilbert 1890). Slope processes were, however, frequent in the silt sediments below the sand. The proximal slope deposits of fine grained deltas consist mainly of deposition from the suspended load. This is characteristic of fine grained deltas with lateral flow separation (Axelsson 1967). The abrupt discontinuity in channel profile at the foreset slope of a laboratory sand delta was gradually subdued and finally eliminated, when silt and very fine sand was added to the flume (Jopling 1965).

The sand sequence is predominated by tabular cross-bedded sets. These are generally considered the result of migrating linguoid or transverse bars with foresets, representing avalanche slopes (e.g. Collinson 1970, Smith

1970, 1972, Cant and Walker 1978). The tabular cross-bedded sets and the asymmetrical ripples show unidirectional flow with only small variations within the sequence.

The migration of transverse or linguoid bars is characteristic of distal outwash rivers, and depositional sequences from sandy braided rivers as River S. Saskatchewan (Cant and Walker 1978) and River Platte (Smith 1971) are dominated by tabular cross-bedded sets. The bars are developed in flow expansions in the active major channels.

The ripple cross-lamination and the planar laminated sand show fluctuations in the flow regime. This may be due to changes in flow velocity but also due to changes in water depth. The sand sequence shows a gradual aggradation. Erosional contacts and reactivation surfaces, similar to those described by Collinson (1970), are almost absent. That means that there are no evidences of sand flats emerging during low stages. The channel was probably continuously covered by water. The changes in flow regime, as reflected by the ripples and the plane beds, may, however, represent water depth changes. The channel may represent a distributary on an outwash delta.

The first subaerial indications in the sequence were found in the alternating bed facies on top of the sand. Dessication

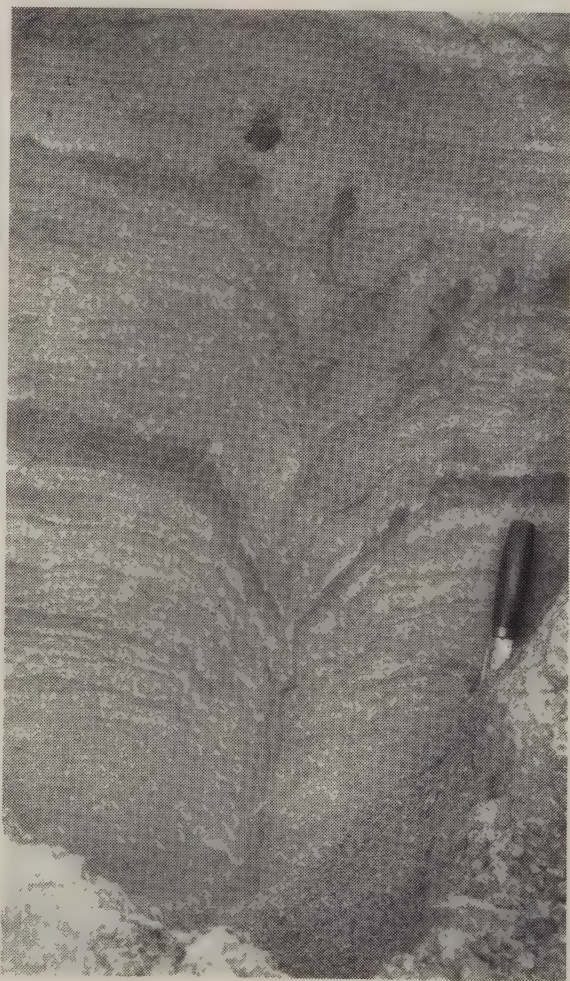


Fig. 88. A syngenetic ice wedge cast in overbank flood deposits.

structures in the clay laminae and ice-wedge casts were formed intermittently in the place. The 30 to 35 fining-upward sequences show structural changes, similar to the Bouma sequence with planar-laminated sand, A-type ripple lamination, B-type ripple lamination and draped lamination of silt and clay. The sequence was deposited during one single flow and is interpreted as being formed during short lived overbank flooding in distal part of the outwash floodplain. The convoluted silty sand bed between the channel deposits and the overbank deposits is settled out from liquefied flow and may be a crevasse splay or a collapse of a proximal levee.

The transition into a submerged continuously water-covered environment is shown by the silt and clay facies in the upper part of the sequence. The sedimentation in the lake was a slow settle of clay particles, interrupted by underflow currents, which deposited current ripples and parallel-laminated silt. The lake floor was affected by oscillatory water movements, generated by waves. The thickness of the turbidity flow deposits decreases upwards, and at the same time the symmetrical ripples disappear. This must be due to a rise of the lake level, giving an increasing distance to the inflowing river and a position of the lake floor below the wave base.

6.2.2 Environmental interpretation

A reconstruction of the changes in sedimentary environment during the Glumslöv interphase necessitates two assumptions. The first is that the dislocated rafts are of a local origin and that the primary position has been immediately to the east or south-east for the Ven rafts and immediately to the south for the mainland rafts. The sections can thus be arranged in an approximate

transection from Ven eastwards to the mainland and another transection from the southern part of the mainland towards the north. The second assumption is that the sedimentation happened in one large basin, and the sedimentary successions can thus be correlated according to water-level changes, which were simultaneous events in the basin. A pervading character of the environmental changes, interpreted from the lithostratigraphic descriptions, is the changes of the water level. The construction of the correlation chart, (Fig.89), is based on two events, the regression maximum in the lower sand bed of the member, and the transgression maximum in the upper part of the sequence.

The sedimentation during the Ålabodarna phase deglaciation is everywhere connected to subaqueous density flow processes or fall-out from suspension in a glaciolacustrine environment. The subaqueous environment during the Glumslöv interphase was rapidly changing into subaerial or fluvial conditions. This happened approximately after 40 years in the NW part of the area. A braided stream environment, often in combination with ice-wedge casts and scoured surfaces, shows that the basin emerged above the lake level. Only site 1B shows neither signs of interruption in sedimentation nor subaerial delta deposits, and the planar laminated sand in log 2 is chosen as approximately corresponding to the regression maximum, and it was deposited slightly below or close to the lake surface. All other sites were by that time exposed above the lake level.

The emerged area was a flat, silty plain, slightly inclining towards the east. Further to the east the bedrock elevation of the Helsingborg ridge rose above the plain. A fine sand alluvium was spread out in the area. The fluvial character of the lower sand bed in the Glumslöv member is evident. The lateral spreading

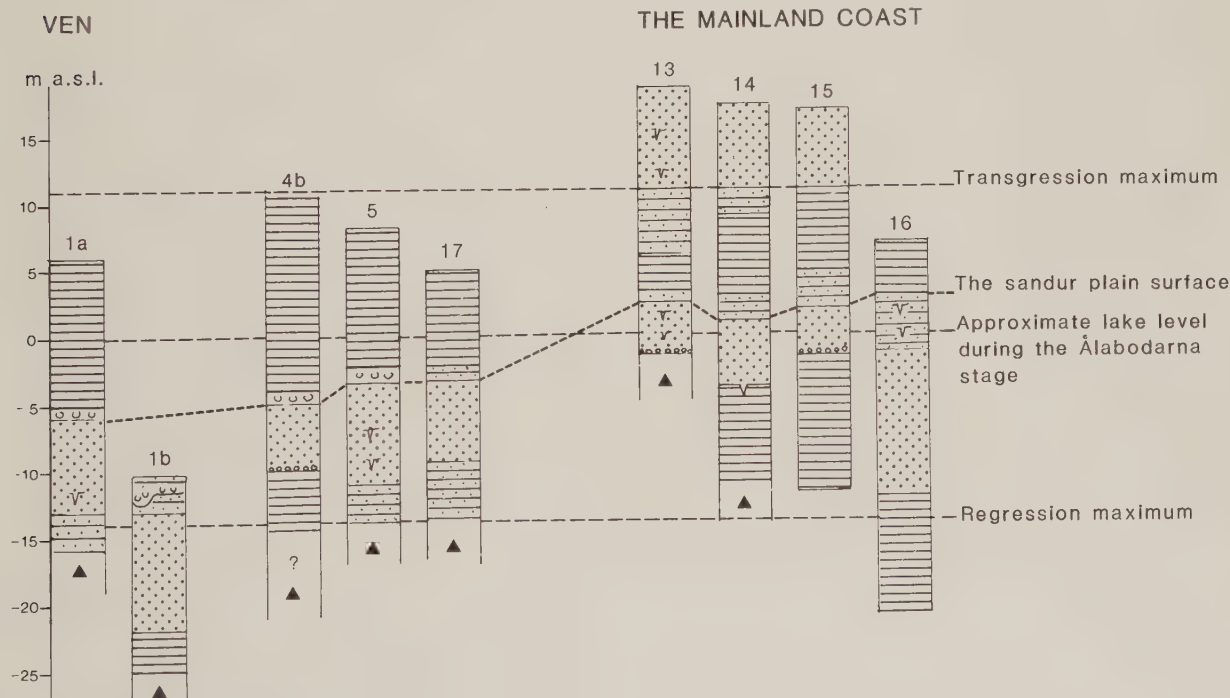


Fig. 89. Correlation chart showing the interred vertical position of the most extensive lithostratigraphic records from Ven and the mainland coast. The amplitude of lake level stages and the top surface of the sandur plane is also shown. The geographic position of the sites is shown in Fig. 5.

(a rough estimation is $>25 \text{ km}^2$), mean sizes in the fine sand fraction, a good sorting and the internal structures of the sand indicate a distal braided stream environment or rather an alluvial plain environment. The average flow direction was from the east or the south-east. The gradient of the braided stream delta can be approximately evaluated from the top of the fluvial sediments in Fig. 92. The distance from the west coast of Ven to the mainland is about 7.5 km. The elevation of the sediments is 9 m, which gives a gradient of 1.2 m/km. This is a very approximate value, but it corresponds very well with gradients from modern braided fluvial sand environments. Boothroyd and Ashley (1975) found gradients less than 2 m/km in the lower fan sand areas of a braided stream. Boothroyd and Nummedal (1978) got the same value from the distal margin of an alluvial fan.

Two sources for the stream water are conceivable. The stream was either glacio-fluvial or it was a river, draining a periglacial area with snow melting during spring and summer thaw. The interpretation is important for the estimation of the ice-front position during the Glumslöv interphase.

The petrography composition for samples in the sand belongs to either AB/BA or C/CD petrography assemblages (Fig. 63). The AB/BA samples are restricted to the lower part of the sand sequence. In some cases they are even restricted to the very first layer of moderately sorted gravel which was deposited after an interruption in the sedimentation. The layer contains numerous intraclasts and is interpreted as the result of fluvial erosion of diamictons from the Ålabodarna member.

The palaeocurrent directions in the lowermost part of the sand sequence are more spread than in the upper part, and directions from north to south-east were recognized.

The C or CD petrography assemblages have silurian shales as characteristic rock type. The percentage share is high, with a maximum of 28 % at site 16. The nearest bedrock source of silurian shale eastwards is the bedrock elevation of the Helsingborg ridge. A combination of C and D rock types in a sediment of glacial origin indicates a most probable transport direction from south or south-east. The petrography composition does not exclude a glacial origin.

There is no coarsening upwards in the sand sequence, indicating a decreasing distance to an advancing glacier, and the delta sand is not overlaid by ice contact sediments. If there was a glacier, its position would probably be south-east of the area and at a far distance. The most probable interpretation, according to the petrography composition and the palaeoflow direction from the east, is a periglacial fluvial stream which eroded in older sediments and particularly in the silurian shale bedrock on the slope of the Helsingborg ridge immediately to the east of the delta plain. The periglacial environment is determined by numerous observations of ice-wedge casts in the sand.

It is difficult to estimate how long the delta existed. The investigated sand sequences generally show a rapid sedimentation rate, and none of the sections contain complex, repeated facies sequences from the braided stream environment. The growth of the delta was more a lateral spreading

over the flat area. As the sand sequence is present in all the sections, it is likely that a large area of at least 25 km^2 was covered by the sandur.

The sediments at site 16 in the south-eastern part of the area were abandoned by a channel probably in an early phase of the delta growth and was thereafter only affected by repeated channel overflows. Cryoturbation and ice-wedge casts indicate an annual cyclicity in the sedimentation. 30 to 35 such annual floodplain overflows are registered in the sequence before the site was permanently covered by lake water. 30 to 35 years can thus be regarded as a minimum time for the existence of the delta plain.

The braided delta sand is everywhere overlaid by alternating bed facies and silt and clay facies. The plain was flooded, and the basin passed over into a new lake phase. The deltafront rapidly migrated eastwards, and only very thin subaqueous distal delta deposits lie between the subaerial delta sand and the stillwater sediments.

The sedimentation in the lake was affected by fall-out from suspension, sliding of lake sediments, gravity flow processes and wave action. Distinct clay laminae can be separated from silt layers. The influence of gravity movements and the effect of wave action of the lake floor prevented, however, a development of distinct varves. It is thus difficult to evaluate the time span during which the lake existed. A rough estimation was made at log 1 where the 10.5 m thick silt and clay sequence was interpreted as deposited during 160 to 240 years.

The sedimentary conditions, and wave activity remained more or less constant during the aggradation of the thick sequence. In spite of the high aggradation it seems as if the water depth remained almost constant. If so, a more or less continuous transgression happened during probably more than 100 years.

The transgression maximum is evaluated from sites 13, 14 and 15 where the parallel-laminated lake sediments grade into subaerial delta sand. The delta facies is missing on top of the silt and clay facies on Ven, and the lake sedimentation probably continued in this part of the basin during the aggradation of the delta on the mainland. The top of the lake sediments on Ven is correlated to the base of the delta sand on the mainland in the correlation chart Fig. 92. The transgression can be estimated to at least 25 m.

The rise of the water level in a basin is generally interpreted as the result of changes in the pass point area. This can be due to differences in uplift or subsidence between the basin and the pass point area, or it can be the result of damming by ice in the passpoint area.

The transgression maximum marks the end of the sedimentation during the Glumslöv interphase. As the contact to the next member is discordant and erosive, it is impossible to follow a gradual transition into another sedimentary environment. What happens next is the gigantic dislocation of the lower members by an overriding glacier. The sediments in the topmost part of the Glumslöv member do not indicate an advancing glacier. The fine sand delta at sites 13, 14 and 15 and distal fluvial of the same type as the lower delta, and the lake sediments in the Ven area do not change upwards.

There are two possibilities to interpret

the sudden cease in sedimentation:

a) The glacier entered the lake basin and dislocated the sediments. Proximal glacial sediments (if there were any) were eroded and removed from the area.

b) The lake level suddenly fell, and the area was converted into subaerial conditions without sedimentation. After that the area was invaded by a glacier from south-east. The sudden drop of the lake level was due to changes in the pass-point area or a change of the damming effect from a glacier probably in the Kattegatt area.

6.3 Möllebäcken glacitectonics

6.3.1 Previous investigations

The dislocations of the Pleistocene sediments in the area have been described by numerous authors (Oerstedt 1844, Torell 1872, Erdmann 1873, 1881, 1883, Holmström 1874, Madsen 1917, Hansen 1940, Wennberg 1949, Johnsson 1956, 1962, Markgren 1961 and Rasmussen 1973a, 1973b, 1974). As early as 1872 Torell interpreted the dislocations as made by a glacial agency, and ever since this has been considered the cause of the large deformation structures.

Scattered measurements of fold axes and fault planes were made by Johnsson (1956, 1962) and Markgren (1961). According to these authors, the deformation indicated ice pressure from the south, and the tectonics was connected to the Low Baltic Ice. Hansen had earlier suggested a deformation by the previous NE Ice (1940). His assumption was based on the stratigraphic observation that undisturbed 'NE'-deposits rested on top of vertically erected sediments at Ålabodarna on the mainland.

A systematic investigation of tectonic structures was made by Rasmussen (1973). His work comprises measurements of the tectonics in the cliffs of Ven and the nearby mainland. He found three main ice-pressure directions, a NE and an ESE direction on Ven and a SE direction on the mainland. His stratigraphic investigations were very brief, and the glacitectonic events were correlated to the traditional ice-stream model. The two glacitectonic directions on Ven were connected to the NE Ice, and the tectonics on the mainland was connected to the Low Baltic Ice (Rasmussen 1973a, 1973b).

A normal fault with 11 m displacement at Sundvik clay-pit was described and discussed by Madsen (1917). The N 40°W strike did not fit into the general glacitectonic pattern with ice-pressure directions from south or south-east. The strike was the same as the trend of the bedrock lineaments in Skåne (see Fig. 2), and he connected the normal fault to bedrock movements during the last glaciation.

6.3.2 General considerations on glacitectonics

Exodiamict glacitectonic deformation (nomenclature after Banham 1975, 1977) is the result of shear stress on the substratum, imposed by an overriding glacier. Glacier ice is very weak, and shear stresses in glaciers seldom exceed 1.5 bars (Paterson 1969). An assumption for the glacitectonic deformation is that the shear strength of the substratum is less than the shear stresses, imposed by the glacier.

Unconsolidated sediments are a two-phase system where the interstitial water is the factor, most affecting the physical properties (Hills 1972). Sediments of different granularity and permeability react in different ways if the pores contain water. They will also react differently depending on the pore-pressure, relative to the external confining pressure. The shear strength is a function of the effective normal stress, which is defined as the total normal stress minus the pore-water pressure (Tersaghi and Pech 1976). If the pore-water pressure is zero, a total normal stress, representing 35 m of ice thickness on sand and 70 m on clay, is enough to prevent deformation (Boulton 1972a). The elevation of pore-water pressures is crucial for the subglacial deformation. The strength of unconsolidated sediments is increased if the sediments become frozen, and the strength continues to increase with a further decrease in temperature (Muller 1947), and frozen sediments are unlikely to be deformed (Boulton 1972).

Deformation during high effective pressure is largely by cataclasis. When the pore-pressure and the confining pressure are equal, the sediment is weak, the internal friction is reduced, and the deformation is by intergranular movements (Hills 1972).

6.3.3 Deformation characteristics and interpretations

The tectonics of the cliffs comprises both folding and faulting. The relationship between faults and folds in the glacitectonic deformation must be considered. The deformation may start as a plastic deformation of a ductile sediment and will, with increasing effective pressure, reach the point of rupture and continue as a thrust deformation. The deformation may also consist of thrusts with drag folds, developed during the thrust movement. This distinction is important for the understanding of the glacitectonic deformation in the investigated area.

The stratigraphic position of the Möllebäcken glacitectonics is between group II and group III strata (Fig. 62), and the youngest sediments, affected by the tectonics, are of the Glumslöv member. The lower boundary of the deformation lies below the present sea level. The Uranienborg member was not found in the cliffs, and it is likely that the lower limit of the deformation approximately coincides with the boundary between the Ålabodarna member and the Uranienborg member.

The shear stress directions, belonging to the Möllebäcken glacitectonics, are from E or ESE on Ven and SE to SW on the mainland coast (Fig. 91).

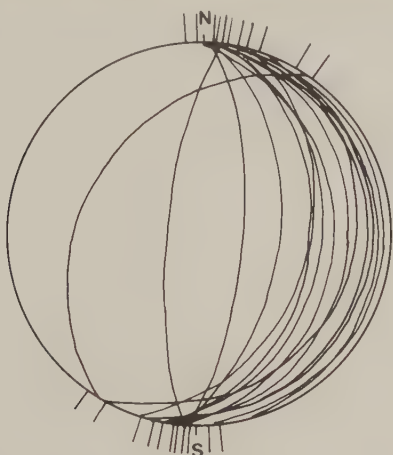
The type of deformation is best interpreted from the large exposures at site 1 and site 4. The strata at the two sites are bent in large basin-like flexures. The general dip of the layers in the western part is towards the east and in the eastern part towards the west. The boundary between the Ålabodarna member and the Glumslöv member at site 1 lies at 25 m above the sea level in the north-west part and dives down below the sea level in the middle of the section and is found again at a level of 6 m above the sea level in the south-eastern part. The same boundary in section 4 lies 20 m above the sea level in the western part, dives below the sea level in the middle part and lies at about 17 m above the sea level in the eastern part.

The flexures are transversed by series of reverse faults. The dip of the fault planes



Fig. 90. Möllebäcken glaciotectionics. Inclined strata of the western limb of a synclinal bend at site.1.

Ven



The mainland

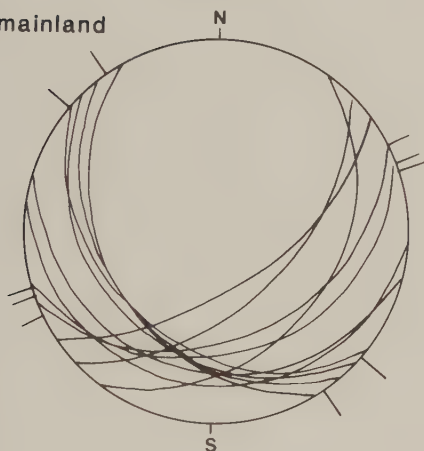


Fig. 91. A complication of measurements of the Möllebäcken glaciotectionics. Stereographic projection. Thrust faults are plotted as large circles. The trend of fold axes is shown by marks outside the circle.

in section 4 changes from the east to the west. The fault planes have a rather low angle of dip in the eastern part but steepen towards the west. Some almost vertical thrusts are to be found in the western part. None of the shear planes have been developed along a clay or silt lamina, and the faults intersect the stratification obliquely. The internal deformation of the individual thrust blocks is generally very small. Sometimes drag flexures were developed in the sediments, adjacent to the thrust planes during the shear movement.

The increased angle of dip in the ice-flow direction is opposite of what has been described by Slater (1927) as a common feature in glacially dislocated sediments. Köster (1956) has also shown by experiments that imbricated thrust blocks in front of a glacier model show progressive steepening of faults towards the glacier.

The increasing inclination of the shear-planes in section 4 can, however, be explained. The deformation has started as a synclinal bend with a north-south axis, and then the eastern limb has been faulted in a series of low-angle overthrusts. The increased influence of the resistance of the western limb has resulted in a steepening of the shear-planes.

A conceptual model for the Möllebäcken glaciotectionics is shown in Fig. 92.

6.3.4 Environmental interpretation

The deformation of the sediments was mainly by cataclasis and the weakest sediments, from where the deformation started, must be found below the Ålabodarna member. The stress was consequently transmitted through more than 30 m of mainly silt and clay sediments. The shear strength in the upper part must then

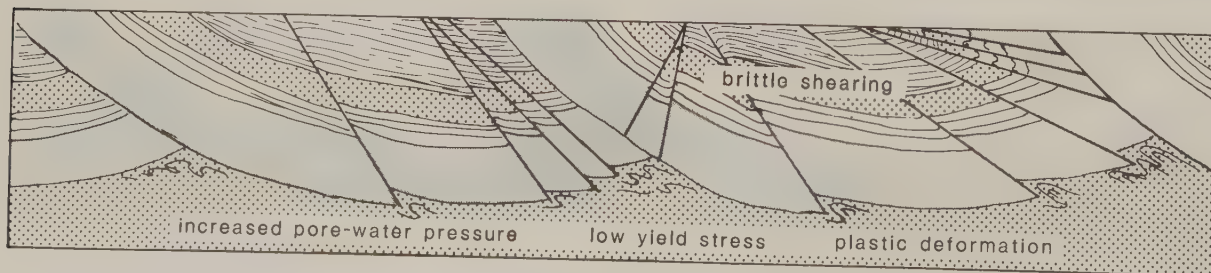


Fig. 92. Conceptual model for the Möllebäcken glaciotectionics. The strata comprise Uranieborg member, Ålabodarna member and Glumslöv member.

have been relatively high, while the strength in the substratum was minimal.

The interpretation of the environmental conditions during the late stage of the Glumslöv interphase was a glacial lake with a delta. The shear stress, required to deform water-saturated clays in general, is very low, and the strength of the glacial lake sediments was probably increased, either by increased normal stress from the glacier overburden or by permafrost. Permafrost conditions were interpreted from ice-wedge casts in the delta sand in the eastern part of the area.

The reduced strength of the sediments below the Ålabodarna member was obtained by the elevation of pore-water pressure in the Uranienborg member. These mainly sandy sediments can be connected to the extensive sand deposits in the Alnarp Valley, which are one of the largest aquifers in the south of Sweden (Leander 1971, K. Nilsson 1971). The aquifer was confined either by permafrost or by the impermeable layers on top of the sand, and it was possible to get an increased pore-water pressure in the sand. The importance of the uplift forces in buried aquifers during glacial tectonism was emphasized by Clayton and Moran (1974) and Moran et al. (1980). Thus, the specific hydrological conditions in the Alnarp Valley did not necessarily prerequisite permafrost conditions, even if the entrainment and transport of large erratics are often explained by the freezing on of a permafrozen substratum to a cold-based glacier (Boulton 1972).

The topography of the landscape is another important factor in the development of glaci-tectonics. Valley sides and islands are generally more susceptible to glaciectonic deformation than flat areas (Banham 1975).

At the time before the glacier advance the landscape was a flat delta plain and glacial lake area with a distinct bedrock elevation to the north-east. This gives good prerequisites for compressive flow, when the glacier reached the bedrock elevation.

The overthrusts were stacked from south-east and south-west on the mainland and from the east on Ven. The divergent glaciectonic directions can be explained by a lobe-formed glacier margin.

The lithostratigraphy in the Ven-Glumslöv area shows no deposition in connection to the glacier, which dislocated the sediments during the Möllebäcken glaciectonic event. This can be interpreted in two ways. 1) A till was deposited on top of the dislocated sediments but later completely eroded. 2) The glacier was cold-based and did not deposit any subglacial sediments. The absence of sediments favours but does not prove the later interpretation.

6.4 Västernäs member: the deglaciation of a temperate, active glacier.

The distribution of facies within the area is shown in Fig. 62. The E1 facies has the largest lateral extension and can be found as a discontinuous layer in the whole area. Sand, gravel and E3 facies are generally found as laterally restricted beds (sites 1, 3, 4, 6, 14 and 15) with the exception of the sections on the north-coast of Ven (sites 7, 8, 9 and 10). The strata in the northern cliffs are quite well exposed, but still they provide difficulties in the lateral tracing and stratigraphic correlation. This is partly

due to the glaciectonic deformation, which has affected the lower part of the member (group 11 strata in Fig. 61), but it is also due to a complex pattern of relationship between laterally restricted beds of different facies.

The boundary between the Västernäs member and the Laebrink member is in most cases erosive, and it is impossible to interpret how much of the lower member was eroded. The present pattern with complex and thick strata, restricted to a small part of the area, and a general, widespread distribution of the E1 facies are, however, considered approximately representing the original distribution of the various facies.

The fissile diamicton facies

Description

The homogeneous diamicton with fissile structures was found at site 2, 4, 8, 10 and 15.

The average grain size composition of the homogeneous diamictons is shown in Table 4, p.52. The total silt and clay content is approximately 50%. The primary mood falls within 2-4 ϕ size grades. The boulder content in the diamictons is moderate or low, but compared to diamictons from the other members, the frequency is high (Fig. 93). Striated clasts are frequent.

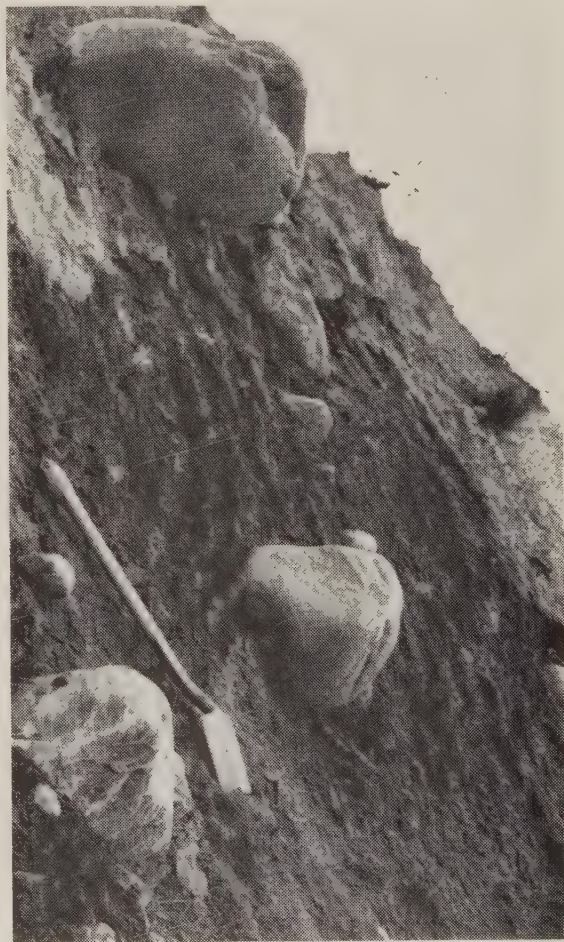


Fig. 93. The boulder content in the homogeneous diamicton facies of the Västernäs till member (from site 4).

The fissility pattern was studied at sites 2 and 4. It consists of subhorizontal undulating fissures, which conjugate into flat-lying, irregular lenses (Fig. 94). Weakly developed slickensides were recognized by stroking with the finger in different directions. The surface felt smooth, when stroked from the north-east, and rough in the opposite direction. The maximum thickness of the lenses generally ranges between 2 and 10 cm. The fissure planes diverge around cobbles, and the lens thickness is often determined by the diameter of the enclosed cobbles. A secondary fissure pattern of high-angle or vertical fissures is developed. Some fissures are curvilinear and meet a lower horizontal fissure-plane tangentially and the upper horizontal fissure-plane at high angles. These fissures are generally restricted to a single lens, while other rectilinear fissures sometimes extend over several lenses across the horizontal fissure planes.

Fabric analyses at sites 2 and 4 show a preferred orientation of elongated clasts. Clasts in the preferred orientation have a general dip of 25° to 40° towards the north-east. The analyses also show a weak transverse trend with almost horizontal A-axes (Fig. 95).

Depositional processes

The diamicton contains a completely new assemblage of rock types, compared to the substra-

tum (see chapter 5.2.3.). The grain size composition, with particles from clay size grades to boulders, sometimes more than 1 m in diameter, indicates a non-sorting transport medium of high competency.

The fissility pattern and the orientation of clasts were produced under the influence of shear stress. A distinct preferred orientation is developed during intergranular strain deformation. The irregular fissility pattern indicates a minimum of movement along shear-planes during the final stage of shear deformation. The orientation of clasts and the fissility pattern is interpreted as formed in a traction zone of deforming bed (Boulton 1975), which prerequisites a high pore-water pressure during the deformation.

The textural and petrographical homogeneity of the diamicton bed can also be explained by an efficient mixing in the subglacial zone of traction (Eyles and Menzie in Eyles 1983).

The combination of properties in the fissile diamicton of the Västernäs member points at a glacial transport and a subglacial deposition under a dynamically active glacier with a base at the pressure melting point. The fissile diamicton displays many of the characteristics, commonly used to recognize a lodgement till (INQUA Commission on Genesis and Lithology of Quaternary Deposits, Work group 1, circular no. 18).

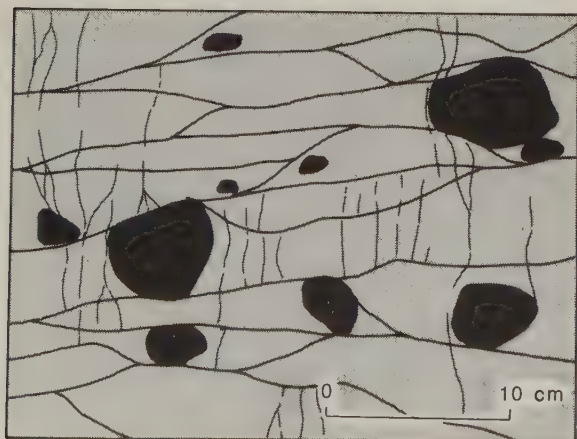
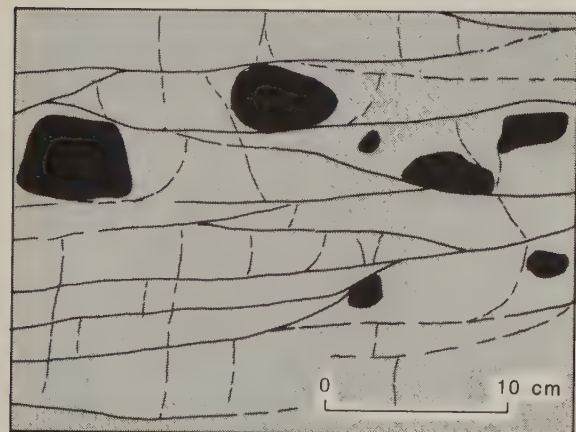


Fig. 94. Fissility pattern in the fissile diamicton facies. A. Fissility pattern parallel to the preferred orientation. B. Fissility pattern perpendicular to the preferred orientation.

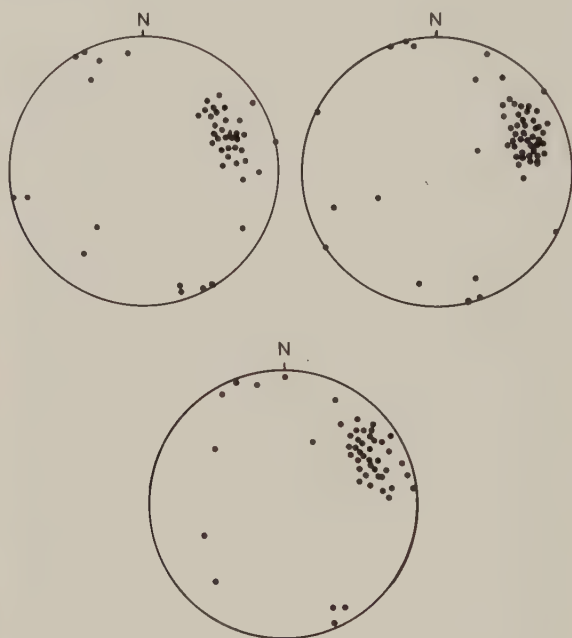


Fig. 95. Clast long-axis orientation and dip in the fissile diamicton facies, Västernäs till member.

The facies complex at Västernäs

Description

The strata of the Västernäs member in section 8 can roughly be divided into a lower diamicton unit (8a:1, 8b:1), stratified diamicton, sand and gravel (units 8a:2, 8b:2) and an upper diamicton (unit 8a:3). The lower diamicton is, however, not always a basal bed in the sequence, but in some places it is stacked together with the stratified sediments and even overlayers these sediments (unit 8b:3).

Evidences of deformation are frequent (see Fig. 29). The stratigraphic position of the deformation is, however, problematic. Part of the deformation is connected to the lower boundary of unit 8a:3. This deformation level can also be recognized at site 9. Part of the deformation did, however, exist during the deposition of the stratified sediments (units 8a:2, 8b:2), as these sediments are restricted to shallow synformal depressions in the substratum. Two events of deformation can thus be recognized in the sediments. The first deformation is connected to the formation of the lower diamicton, and the second deformation is connected to the lower boundary of the upper diamicton.

The lower diamicton (unit 8a:1, 8b:1) was described as a homogeneous diamicton with a basal fissile facies, which was seen only in the eastern part of the section, and a homogeneous diamicton with sand partings, which was traced along the entire section (see chapter 5.1.8).

The fissile diamicton is of the same type as has already been described.

The non-fissile upper part of the unit can be divided into distinct layers, separated by sand partings (Fig. 97). The sand laminae are quite continuous, and in the parts of the section, where the deformation is least comprehensive, the sand streaks can be traced for tens of meters. The sand streaks are very thin, generally less than 0.1 cm. In a few places the laminae increase in thickness. The increased thickness has the form of lens-shaped expansions of the continuous sand laminae. The sand partings and the lenses are in some places deformed into overturned hook-formed folds (Fig. 97b). Slickensides with minute steps, generally less than 0.1 cm high, are developed in the thin sand partings. In cross sections the structures are formed as series of parallel overthrusts with a displacement of 0.1 to 0.3 cm. A horizontal preparation of the sand laminae shows parallel micro-ridges. The orientation of the linear ridges is N 15°W and N 55°W, measured in two different parts of the section. In some places the sand partings are bent over a cobble or a boulder (Fig. 97c).

The diamicton layers between the sand partings range in thickness between 8 and 15 cm with an average thickness of approximately 10 cm.

Three fabric analyses were made in the homogeneous diamicton with sand partings. All the analyses show a preferred orientation with a NW trend (Fig. 96). The direction of a-axes diverges approximately 90° from the striae directions on boulders in the diamicton (Fig. 96). There is also a striking difference in the orientation of the striated boulders, compared to the orientation of the pebbles in the fabric analyses. Most of the boulders have their longaxes perpendicular to the longaxes of the pebbles, and stoss-sides of the boulders are directed towards NE.

Striated and flat-iron shaped clasts are present in the diamicton with sand partings.

The stratified sediments, which partly overlayer and partly are stacked together with the homogeneous diamicton, are laterally restricted. They are concentrated to shallow synformal depressions in the diamicton surface. Well sorted sand and sandy gravel predominate, but layers of poorly sorted gravel and diamicton layers are also present. The diamicton beds are most frequent in the outer part of the synform infillings, where they interfinger with the sand and gravel. The thickness of the diamicton beds ranges from a few centimetres to 0.5 m.

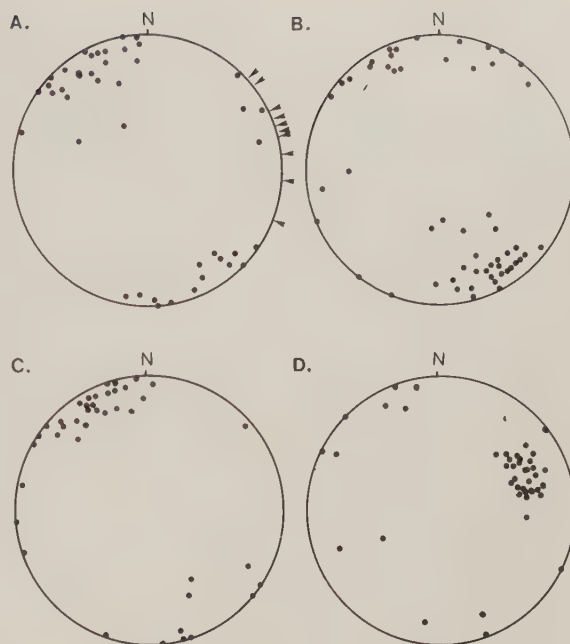


Fig. 96. Clast long-axis orientation and dip in the diamicton facies with partings (projections A, B and C). Striae directions on boulders are marked by arrows outside the circle in projection A. Fabric analysis from the fissile diamicton (projection D). Section 8.

The layers are sometimes wedge-formed and sometimes irregular with erosive contacts to adjacent sand or gravel beds.

Most of the diamicton beds in the stratified facies have a sandy matrix.

The thick sand sequence at 8:120-150 lies in a depression with sharp lateral contacts to the homogeneous diamicton. The sand sequence shows a coarsening upward trend with cross-laminated and planar-laminated sand in the lower part and gravelly sand with planar lamination and trough cross-bedding in the upper part.

Interpretation

The textural and petrographical composition of the diamicton layers is very homogeneous and similar to the composition of the fissile diamicton, which was interpreted as a lodgement till. Striated and flat-iron shaped clasts indicate a glacial transport in the zone of traction prior to the deposition (Boulton 1978).

The ice-flow direction, as interpreted from the lodgement till, was from NE. The orientation of the clasts in the non-fissile facies is perpendicular to the inferred ice-flow direction. A perpendicular orientation of clasts is expected to develop in zones of compressive flow in a glacier (Boulton 1970). The preservation of a perpendicular orientation during deposition beneath an active glacier is not likely. The lodging process or the shearing of debris, melted out from the glacier, reorients the clasts in a direction, parallel to the ice-flow direction.

The transverse orientation of clasts in the present diamicton is very consistent. This shows that shear strain by intergranular movement did not occur during the final deposition. The orientation of the boulders is,

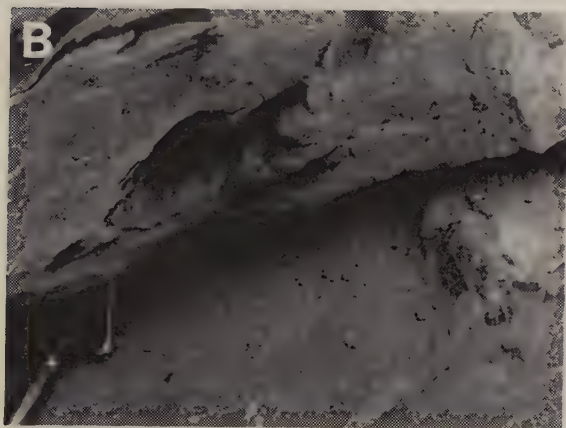


Fig. 97. A. Diamicton facies with sand partings and sand lenses. B. Hook-formed folds with slickenside structures. C. Conform bending of sand partings over a boulder. D. Close-up of the homogeneous diamicton facies with sand partings. From section 8.

however, parallel to the inferred ice-flow direction. The large clasts were probably transported in traction and thus became or retained a parallel orientation during the last stage of the transport. The striations on the boulders were then formed by the bypassing smaller clasts, transported in suspension in the ice.

The absence of a fissility pattern and the preservation of a clast orientation, obtained during transport in a compressive flow, are in advantage for a basal melt-out process beneath a stagnant glacier ice.

The formation of the consistent pattern of sand streaks in the diamicton was formed either prior to the melt-out process or contemporaneously to the final deposition. The sharp contact to the adjacent diamicton interbeds indicates a formation process, separated from the till formation. The process of sand sorting was extremely restricted in vertical direction. Only in a few places the sand partings grow into more extensive lenses. The association to the diamicton indicates a basal or subglacial environment, and the vertical aggradation was most likely prevented by overlying ice. The sand grains moved in a film either between the ice and the substratum or along horizontal thin cavities or shear planes within the ice.

Two conceptual models of the formation of the sand partings are suggested.

The first is that the partings were formed prior to the stagnation of the ice and then preserved during the melt-out process. The presence of a film of water in the contact between moving ice and a rock substratum was theoretically discussed by Weertman (1964) and actually observed by Vivian (1970) beneath the Glacier d'Argentière in France. The formation of a waterfilm must also be possible along shear-planes in the glacier, where the impermeable ice prevents infiltration of the water, produced by frictional heat during the shear movement. Shear-planes are formed in zones of compressive flow when the strain rate is not sufficiently rapid to release the stresses within the ice. This is likely to occur in the thin ice near the glacier margin, where the creep rate, according to Glen's law (Glen 1955), is low.

The jig-saw structures along the sand partings could then be related to the shearing along these shear-planes.

The second model is connected to water escape during the melt-out process. Such subglacial and englacial drainage systems in stagnant ice have been discussed and described by Shaw (1979) and Haldorsen and Shaw (1982) and can be used as a criterion of the recognition of melt-out till. In this case, the water flows in cavities within the stagnant ice or at the interface between the ice and the substratum.

The non-uniform subsidence with draping of layers over underlying clasts indicates that the banding was present in the ice before melting (Shaw 1979). The stacking of overlapping plates, consisting of diamicton with sand partings, was prior to the Västernäs glaci-tectonic event, and the basins with glaciofluvial and till flow sedimentation existed before the second glacial advance (see below). The sand partings were cut off during the stacking and were evidently formed prior to the first deformation. Thus they cannot be compared to the stratified sand lenses, described by Shaw (1979) as having been formed in subhorizontal cavities during the melt-out process.

The sorted sediments on top of the melt-out till were deposited by current water. The

interfingering diamicton layers were formed by debris flow. The till flows are generally restricted to the outer margins of the small depressions, and it is evident that the basins existed during the sedimentation of the stratified sediments. Debris-rich ice at the edge of the basin delivered till flows, while the current water and the sand sedimentation predominated in the deeper central part of the basins.

The sand and sandy gravel at 8:120-150 was deposited in a deeper channel between stagnant ice masses.

The whole sequence represents the sedimentation in a supraglacial landsystem (Paul in Eyles 1983), which has been explained in a conceptual model by Boulton (1972).

The outwash sediments

The banded diamicton disappears below the beach level at 9:40-50, and the stratified, sorted sediments increase in thickness. The strata in the cliff between site 9 and site 10 are mainly covered, but the sorted sediments can be traced in scattered minor outcrops, and the unit has a lateral extent of approximately 450 m in S 10°E direction.

The strata show a lateral change with coarse gravel at 9:0-40, sandy gravel and gravelly sand in the upper part of the sequence at 9:50-100 and sand as the predominating sediment in section 10:0-120. The sediments at 9:45-100 show coarsening upward sequences with a gradual change from laminated silt and clay to alternating bed facies, sand facies and finally cross-bedded gravelly sand and sandy gravel at the top.

Description of strata

The best exposures in the stratified sorted sediments were found at site 10, where a series of minor profiles were investigated. The sand body at the site has a thickness of at least 10 m. The lower part of the cliff was mainly covered, and the measured logs were concentrated to the upper and middle part of the sand body.

Trough cross-bedded sand facies (Facies Bltr)

Trough cross-bedding occurs either as simple cosets or as composite cosets, associated to ripple cross lamination (B 3) and planar lamination (B 2). Individual sets are between 5 and 100 cm thick, measured in the deepest part of the troughs. The foreset beds in logs 1, 2 and 3 are not graded, and gravel clasts are dispersed in the foreset laminae. The foreset laminae in logs 6, 7 and 12 are graded with gravel clasts, mainly intraclasts of clay at the toe of the foreset laminae. Trough cross-bedding is most frequent in the northern and upper middle part of the section. The thickest bedsets are found in the northern part. Fig. 98 shows trough cross-bedding from this part.

Tabular cross-bedded sand facies (Facies Bltab)

Tabular sets of cross-bedding generally occur in composite cosets in combination with planar lamination (B 2) or ripple cross-laminated sand (B 3). More than 80% of the individual sets are less than 25 cm thick. Gravel clasts occur in some of the foresets. The main part of the clasts consists of intraformational clay. The clasts are concentrated at the toe of the foreset laminae. Both angular and tangential floor contacts of the foreset laminae occur. The dip of foresets ranges

between 15 and 35 degrees. Regressive ripples at the base of the foresets are seen in log 5 (180 cm) and in log 10 (140 cm) (Fig. 99a).

Scoured surfaces were not recognized in combination with the tabular sets.

The tabular cross-bedded sets are most frequent in profiles 3 and 4 and in the lower part of the southern profiles. The foreset beds in these parts are thicker than the average. Solitary cross-bedded tabular sets in the upper parts of the southern profiles are thinner than the average.

Planar laminated sand (Facies B2). Sets of planar laminated sand are found in all the sequences and make up a great portion of many of the logs. The individual sets are generally less than 20 cm thick but may be more than 1.5 m thick (log 3, 3.7 to 5.25 m). Planar-laminated sand (B 2) is closely associated to cross-laminated sand (B 3), and facies transitions from B2 to B3 and B3 to B2 are the most frequent in the sequences. Transitions between facies B2 and facies B1tab are also present. The lamination in the planar-laminated sand is generally very distinct and is sometimes accentuated by differences in the mineralogical composition. A dark-coloured banding of heavy minerals is frequent (Fig. 99 b). The planar-laminated sets are generally thinner and more fine-grained in the upper part of the southern profiles than in the northern and lower parts.

Cross-laminated sand facies (Facies B3). Cross-laminated sand is, together with the planar-laminated sand, the most frequent in the sequences. The cross-laminated sand is closely related to planar-laminated sand and laminated silt and clay. Several types of cross-lamination are present. Current types of A-type are the most frequent. Both sets of trough cross-lamination and sets of climbing ripples occur and are about equally represented. Heavy minerals are often trapped in the scoured troughs of the trough cross-laminated sets.

Type B ripples and sinusoidal ripples are less frequent. The three ripple types occur in cosets with gradual transitions from A-type ripples to B-type ripples and finally to sinusoidal ripples.

Laminated silt and clay facies (Facies Dlam).

Thin beds of laminated silt and clay are frequent in some of the sequences. They often occur as graded bedsets with thin laminae of silt, draped over ripple cross-bedded sand, and an upward gradual change into structureless silty clay. The most frequent upward transitions are from facies Dlam to facies B3 or to a scoured surface.

The sedimentary succession, the percentage distribution of the different facies and the frequency of facies transitions within the individual profiles are displayed in Fig. 100.

Two types of wedge casts were found in the section. One type is represented in the cross-bedded sandy part in the north. This type is represented by small, narrow wedges, generally less than 25 cm deep. Four such wedges have been observed. They were developed in tabular cross-bedded sand with planar-lamination or cross-lamination on the top. Two of the wedges were cut by distinct erosive upper boundaries (Fig. 99d).

The second type was found in the upper part of the southern section. The wedges are narrow, but the depth of the wedges can exceed 2 m. The fissures are syngenetic, and the different phases in the formation can be traced to distinct levels in the sediments. These levels are connected to the upper boundary of fining-upward sequences (Fig. 99f).

Interpretation

The distribution of facies within the sand section shows rapid lateral and vertical changes. The northern profiles and the lower parts of the southern profiles consist exclusively

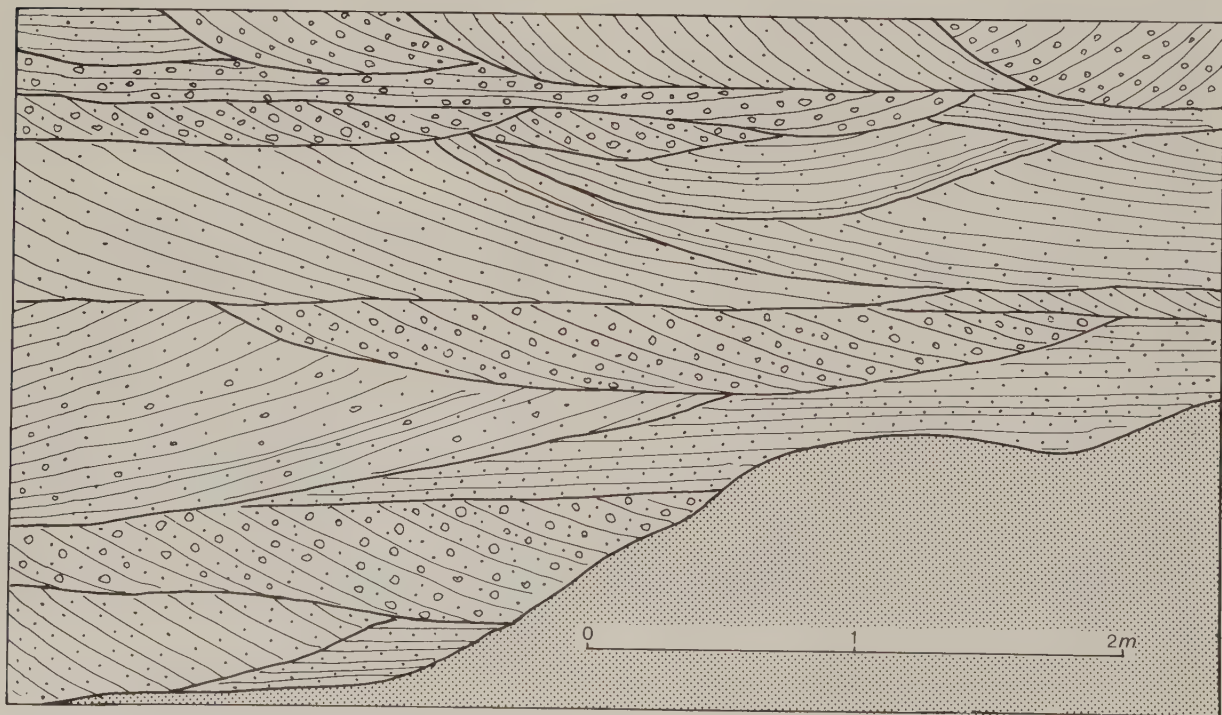


Fig. 98. Trough-formed sets of cross-bedded sand in the northern part of section 10.

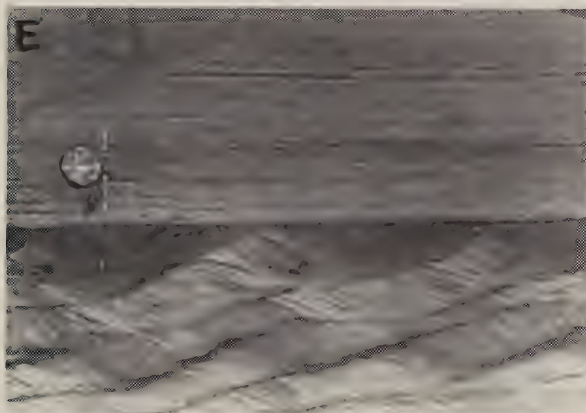
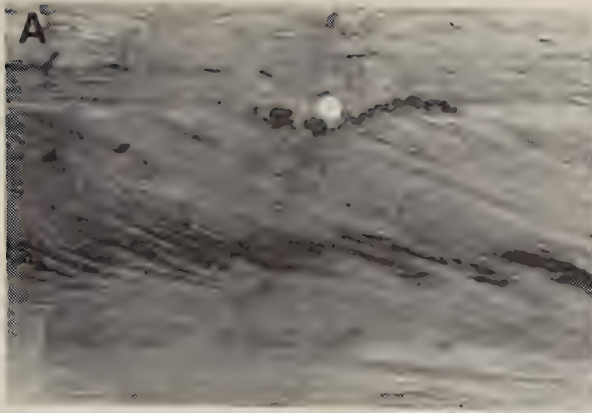
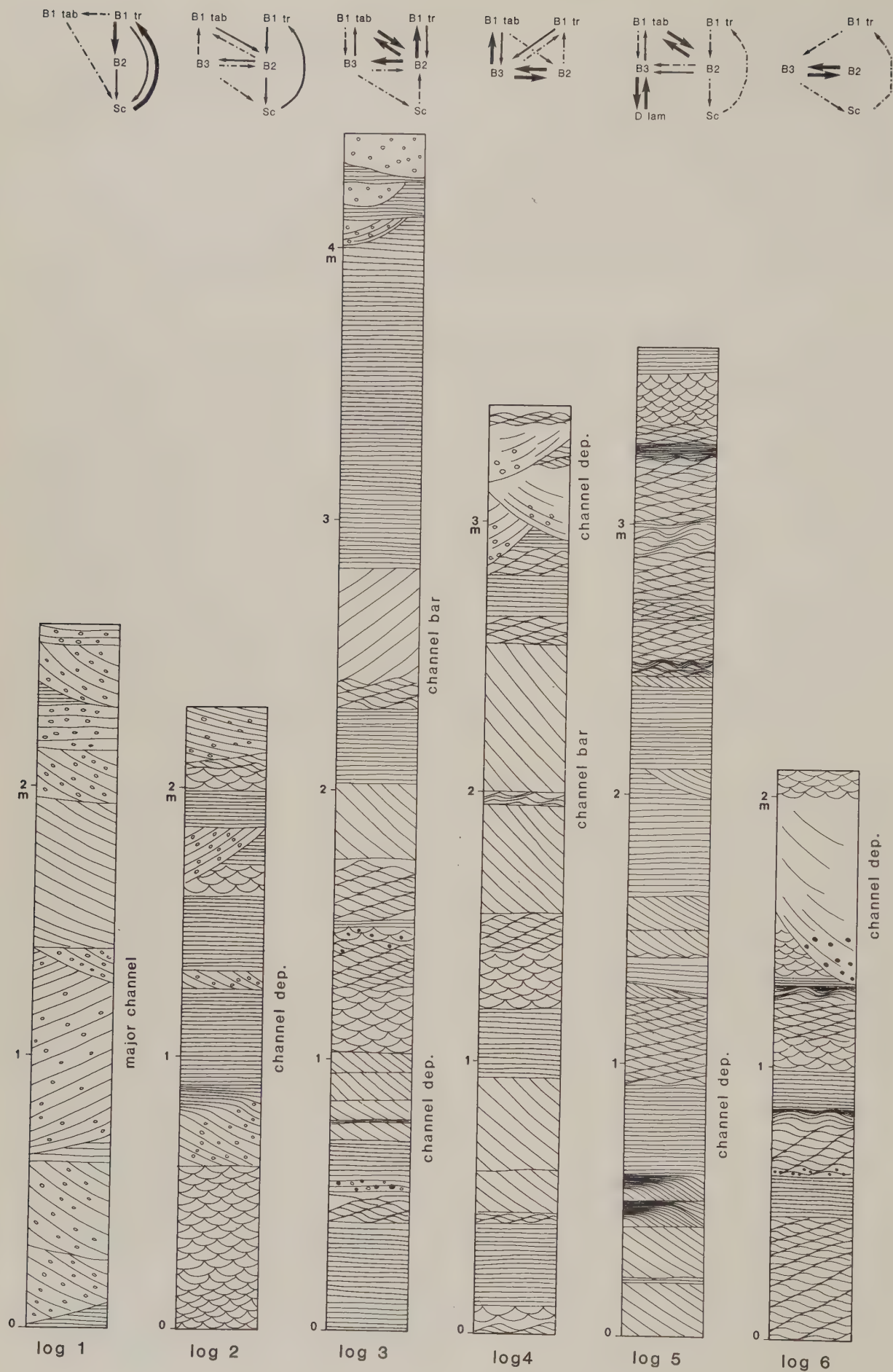


Fig. 99. A. A tabular set of cross-bedded sand with regressive ripples at the toe. B. Trough-formed sets of type A ripples and planar laminated sand. C. Tabular sets of type A ripples grading upwards into type B ripples. D. Soil wedge cast in cross-bedded sand. E. Planar lamination on top of B-type ripples. F. Syngenetic soil wedge cast in overbank deposits.



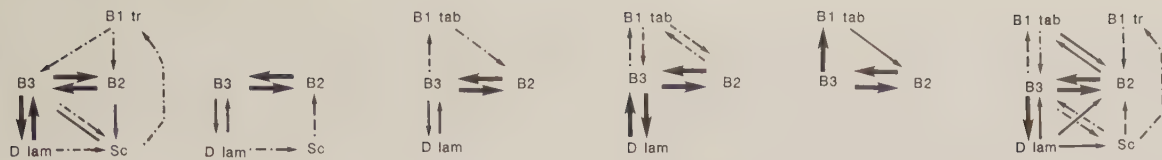
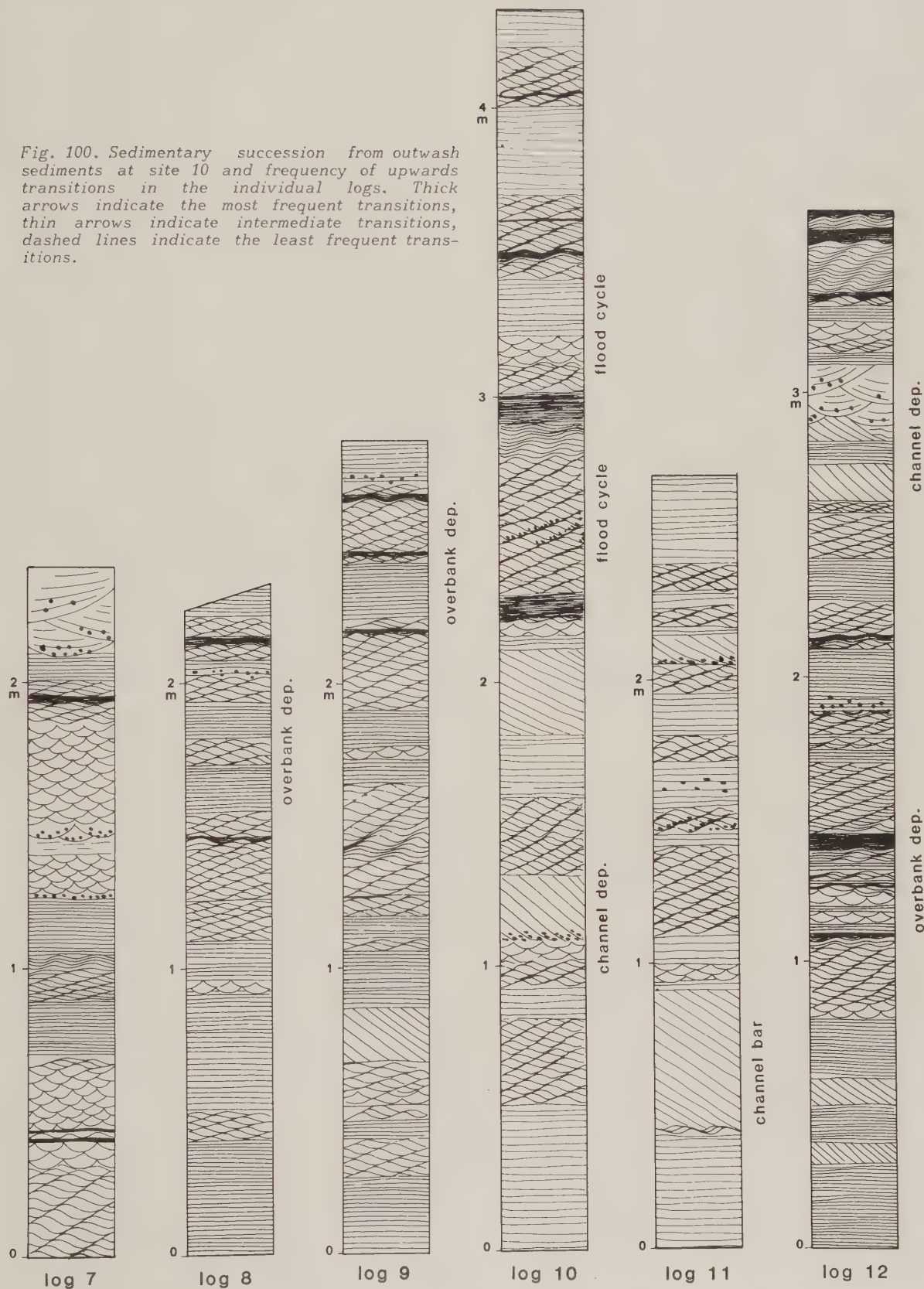


Fig. 100. Sedimentary succession from outwash sediments at site 10 and frequency of upwards transitions in the individual logs. Thick arrows indicate the most frequent transitions, thin arrows indicate intermediate transitions, dashed lines indicate the least frequent transitions.



of bedforms, produced by bedload deposition in unidirectional flow. The profiles from the middle and upper parts in the southern portion of the cliff are dominated by fining-upward sequences, consisting of bedforms from bedload transport in the lower part and draped and parallel-laminated silt and clay from suspension in the upper part. The thickness of the fining-upward sequences ranges between 10 and 100 cm, and the average thickness is approximately 50 cm.

The assemblage of sedimentary structures is typical of a braided river (e.g. Miall 1977).

Trough cross-bedded sand is formed by the migration of dunes (Collinson 1970). The dunes are produced in channels, and their magnitude is directly connected to the water depth (Allen 1968). During flood stages the sizes increase.

The planar cross-bedding is formed by avalanche-slope prograding on linguoid bars (Collinson 1970, Boothroyd and Ashley 1975) or transverse bars (Smith 1970, 1971). The bars are also formed in channels, where the channel curves or widens. Repeated cosets of planar cross-bedding, planar-lamination and cross-lamination are deposited during formation and accretion of sand flats (Cant and Walker 1978).

The fining-upward sequences show rapid, gradual transitions from planar-lamination, deposited in upper flow regime, to mud deposition. Each fining-upward sequence represents an ephemeral stream from overbank flooding. Soil wedges were formed on overbank areas and sand flats during declining and low flow stages.

The different braided stream sub-environments, as interpreted from the profiles, are shown in Fig. 100. The figure also shows the frequency of upward transitions.

The stratified sediments at sites 9 and 10 show a transition from a very coarse proximal facies at 9:0-40 to a distal braided outwash facies at site 10. This gradual transition is characteristic of braided outwash rivers. The short distance from the proximal part to the distal part suggests a small discharge magnitude (Miall 1983).

Västernäs glacitectonics

The second major glacitectonic event in the investigated area is connected to a stratigraphic position in the Västernäs member between group I and group II strata (Fig. 62). Shear stress directions in the Västernäs glacitectonics are from north and north-east (Fig. 101).

Description

The character of the deformation can be best studied in sections 1, 7, 8 and 10. The lower boundary of the deformation at site 1 is a clay lamina. The deformation consists of large recumbent folds with interlimb angles of 0° . The folds are formed in a heterogeneous material of sand, silt and clay layers. The silt and clay layers are plastically deformed and show intergranular strain deformation. The sand layers are fractured. The different ways of reaction on the stress are especially well displayed in the inner arcs of the folds where the sand is intersected by a mosaic pattern of faults, and clay and silt laminae are very much distorted.

The strata at site 7 are also heterogeneous with diamicton, sand, silt and clay layers

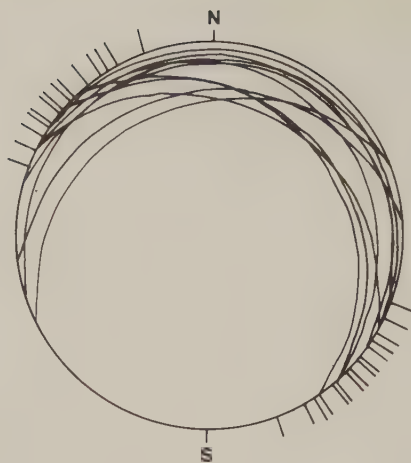


Fig. 101. A compilation of measurements of the Västernäs glacitectonics. Stereographic projection. Thrust faults are plotted as large circles. The trend of fold-axes is shown by marks outside the circle.

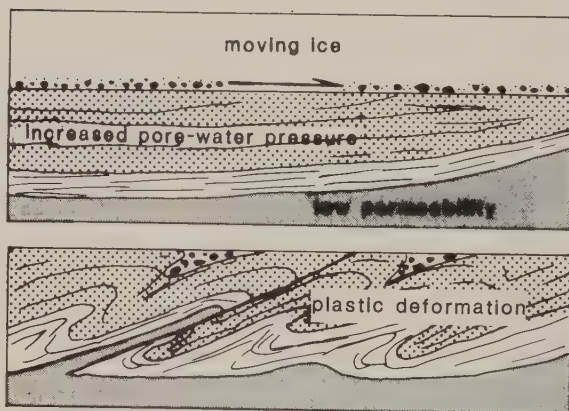


Fig. 102. Conceptual model for the Västernäs glacitectonics. The model is based on the deformation of strata at site 7.

involved in the deformation. The deformation consists of a series of gently inclined or recumbent folds, in some cases transected by thrust faults. The lower boundary of the deformation cannot be seen directly in the cliff, but the lowermost unit, affected by the deformation, seems to be the homogeneous diamicton (unit 7:1), whose upper boundary approximately coincides with the present sea level.

The deformation at site 10 has another character. The sediments consist only of sand, and the tectonics is a series of low-angle reverse faults.

A conceptual model for the Västernäs glacitectonics is shown in Fig. 102. The model is based on data from section 7.

Interpretation

The response to the constant-stress conditions, exerted by the glacier, varies. The competence differences between strata of different lithology can, above all, be seen in the inner arcs of the folds, where the strong compressive stress has resulted in complex flexural-slip thrusts in the sand layers, while the silt and clay layers have reacted as a plastic

material with fold formation.

The same difference can also be seen on a larger scale, where the presence of silt and clay laminae in the sequences was a prerequisite for a general plastic deformation of the strata. The thrusts at site 7 were formed in a closing stage of the fold formation. The sediments did not remain ductile during the increasing effective pressure, and the thrust faults were formed.

The reduced shear strength can be found within the deformed sediments during the Västernäs glaciectonic event. The reduced pore-pressure and the intergranular strain deformation were concentrated to the fine-grained silt and clay layers, and the reduced strength in the clay laminae played an important part in the formation of shear planes. The deformation was shallow and affected a maximum of 5 to 10 m of the substratum. The increased pore-water pressure, which triggered the deformation, was confined to the sediments just beneath the glacier sole.

The upper diamicton

Description

A diamicton on top of the dislocated strata was found at site 8 (unit 8a:3) and site 9 (unit 9a:3). The diamictons rest discordantly on the folded and thrust strata. According to the petrographical composition, the units belong to the Västernäs member, and boulders at the lower contact of the diamicton at section 8 were striated from NE.

The diamicton was studied in a minor excavation at 9:0-10 (Fig. 103). The contact between the diamicton and the gravelly sand in the substratum is a diffuse, two-parted zone with foliated gravelly sand in the lower part and foliated diamicton and sand in the upper part. The sand and the diamicton in the upper part are folded together in tight Z-formed and hook-formed recumbent folds. The upper, foliated part grades into a massive diamicton with a sandy matrix.

A fabric analysis in the massive part shows a weakly preferred orientation in N 60°E (Fig. 104). The dip of a-axes ranges from horizontal to 50°. The inclination of a-axes in the preferred orientation is between 20° and 50°.

The preferred orientation of clast longaxes is the same as the trend of axes in the folded part of the basal zone.

Interpretation

The transition zone between the cross-bedded gravelly sand in the substratum and the diamicton is an example of penetrative shear deformation (exodiamict glaciectonite, according to Banham 1975, 1977). The sand and gravel, which lie in distinct layers with different grain size composition in the substratum, have been completely mixed, and the gravel clasts have been dispersed in a sandy matrix.

The upper part of the foliated zone has a contribution of diamicton, which was released from the overriding glacier by basal melt-out or subglacially transported in a zone of traction. The two sediment types were mixed by folding and thrusting during the subglacial shearing.

The massive part of the diamicton bed is considered a lodgement till. The weakly preferred orientation may be due to the high content of cobbles and boulders, probably affecting the orientation of minor clasts during the strain deformation. It may also be due to a lodgement process by deposition of



Fig. 103. The upper diamicton of the Västernäs member at site 9.

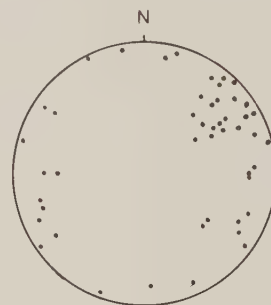


Fig. 104. Clast long-axis orientation and dip in the upper diamicton.

individual clasts or aggregates, which may give a less pronounced preferred orientation. This will be discussed in chapter 6.5.

6.4.2 Environmental interpretation

The ice-flow direction, which was interpreted from the Möllebäcken glaciectonics, was approximately from the south-east sector. The immediate younger strata, resting discordantly on the dislocated older sediments, are from the Västernäs member and have nothing in common with a south-east ice-flow direction. Provenance of rock clasts and the palaeoflow directions in the tills univocally show indications of ice-flow directions from north-east.

The hiatus between the angular unconformity and the lowermost unit of the Västernäs member offers possibilities of speculations. The hiatus may represent a time span with erosion but without deposition or a time span with a combination of erosion and deposition but with a negative deposition balance.

There is no deglaciation sequence from the Möllebäcken glacier, and there are no proglacial sediments connected to a first advance of a Västernäs glacier. This may indicate a continuous subglacial environment from the Möllebäcken phase to the Västernäs phase with changes in the ice-flow direction during a constant ice cover.

The constancy of non-cooperating rock

types in the Västernäs member indicates a rework of material, primarily transported from the south-east sector (see chapter 5.2.3). This might be an indication of an earlier deposition from the Möllebäcken glacier within or to the north-east of the investigated area.

However, a hiatus does not prove anything, and nothing can be stated about the time span between the major deformation from the Möllebäcken glaciectonics and the first deposition of the Västernäs member.

The striking similarities in ice-flow directional indications, grain size composition and petrographical composition of the lodgement till and the melt-out till show a close connection in the formation of the two till facies. It is likely that the lodging preceded the melt-out processes within a short-time interval during the deglaciation.

The deglaciation pattern is shown by the lateral distribution of the sedimentary facies. The main part of the area is covered by lodgement till. In these parts the subglacial land system (Boulton and Paul 1979, Eyles 1983) was revealed during the withdraw of the glacier. The supraglacial land system and the outwash sediments are restricted to the low-lying northern part of the investigated area.

The transverse orientation of clasts in the melt-out till shows a compressive flow in the glacier, preceding the stagnation of the ice. The proportionally thick layer of melt-out till and the deformation structures indicate a stacking of the ice. An application of the supraglacial land system (Boulton and Paul 1979, Eyles 1983) with the deglaciation model of Boulton (1972) very well explains the complex sequences with lodgement till, basal melt-out till, till flows and glaciofluvial sediments in the sections on the northern coast of Ven. Boulton's model was based on observations in Svalbard but can also be applied on temperate glaciers with compressive flow in the outer zone.

The deglaciation indicates subaerial conditions with till flows and glaciofluvial sedimentation in depressions in the stagnant ice. The glaciofluvial sequences at Laebrink (site 7) and at site 9 also comprise glacial lake sediments in the lowermost parts. These glacial lake sediments seem to have a restricted lateral and vertical extent, and it is possible that the silt and clay was deposited in ephemeral ponds, dammed up in the stagnant ice. If the glacial lake had a more extensive distribution, only small shallow bays or creeks would have interfingered into the most low-lying parts of the area. In any case, the glacial lake environment rapidly changed into a subaerial, glaciofluvial environment, and a sandur plain was built up on top of the glacial lake sediments. A braided stream environment was interpreted from the sediments at sites 9 and 10. Several wedge casts within the sediment sequence show that annual soil wedges were formed on sand flats and overbank areas. The plain was restricted in lateral direction towards the south-east by the elevation of the outer raft in the Möllebäcken glacial tectonics.

The glacial readvance that caused the folding and faulting of the strata came from approximately the same direction as the ice-flow during the previous deglaciation. The extent of the glacier margin before the readvance cannot be interpreted from the strata stratigraphy. The oscillating glacier is not registered in the sections on the mainland. The maximum extension of the ice readvance is also unknown. There is no contribution

in rock components or any change in the ice flow pattern that might indicate a separate major advance, and it is likely that the readvance was a minor oscillation during a main ice recession.

The lodging and the subglacial strain deformation and the melt-out processes indicate a glacier with a basal temperature at the pressure melting point. The production of till flows and glaciofluvial sediments shows that the decay of the ice was mainly by melting. The glacier was of a temperate regime during the deglaciation. It is thus likely that the ice recession was connected with a climatic amelioration.

6.5 Laebrink member: units 24-28:

an ice advance across a permafrost area.

Units 24 to 28 in the correlation chart (Fig. 62) comprises the fissile diamicton facies (unit 28) and the strata of the Laebrink member, which underlie this unit. The homogeneous fissile diamicton, which is the marker bed of the member, was laterally traced over long distances, and the unit is almost continuous in the whole area and does not intertongue with other units. The diamicton shows a pervading character of a slight upward change in the petrographical composition and pebble orientation. The D assemblage generally dominates the lower part of the fissile diamicton and is connected to preferred orientations at S10°E-S40°E. The C petrography component increases upwards, and the increasing influence of these rock components coincides with more easterly directions in the preferred orientation of clasts.

The environmental interpretation is based on the assumption that the fissile diamicton is a virtually synchronous unit. The time equivalence is based on the just mentioned criteria, the lateral continuity without intertonguing with other units (Friedman and Sanders 1978, p. 428) and the similar upward changes within the unit.

6.5.1 Sedimentary characteristics and interpretations

The substratum of the fissile diamicton generally consists of strata from the lower members. The main exception comes from the cliff between Fortuna and Ålabodarna on the mainland, where thin but continuous beds of gravel and laminated silt and clay underlie the diamicton and show a sedimentation, prior to the deposition of the fissile diamicton.

Units 24 to 27 at sites 11 and 12 on the mainland coast

Sedimentary characteristics

An extensive layer of very coarse-grained sediments was traced over a distance of at least 300 m in the cliff between Fortuna and Ålabodarna (Fig. 105). The thickness of the layer ranges between 20 and 70 cm, and it lies discordantly on dislocated strata of the Ålabodarna and Glumslöv members. At two sites, 11:110-120 and 12:100-120, the layer is transitional to thicker and more complex strata. The layer



Fig. 105. Coarse sediments overlayers by silt and clay. From the cliff at site 11.

is very heterogeneous and is characterized by rapid, vertical and lateral variations in texture. The texture generally ranges from coarse pebbly sand to poorly sorted gravel with scattered boulders. The gravel is generally clast supported, but small lenses of

a clayey diamicton and lenses of laminated silty sand are in some places frequent in the space between larger clasts. More extensive lenses or layers of better sorted pebbly sand and sandy gravel, mainly with planar lamination, were found in the slight depression between the two sites and at site 12. Boulders were buried in the sediments. The shape of clasts ranges from angular to sub-rounded. Flints are sometimes broken into splinters, and greenstones are partially or completely concentric weathered. A few boulders were striated.

The gravel is overlayers by a 20 to 90 cm thick bedset of laminated silt and clay, and silt and clay also fills up the pore space in the upper part of the coarse gravel layer. The laminated sediments consist of in general 6 to 8 distinct clay laminae, interbedded by parallel-laminated and cross-laminated silt and very fine sand. The thickness of the silt layers generally increases in low-lying parts of the undulating lower surface.

At 11:110-120 and 12:100-120 the coarse, grained layer is transitional to thicker and more complex strata, containing diamicton layers. The petrographical composition of these sediments is similar to the petrographical composition of the stony layer (Figs 43, 44, 47 and 48). The difference is a marked decrease in the content of Cretaceous and Danian limestone and chalk in the coarse, grained layer, compared to the lower diamicton layers.

The stratified diamiction unit was investigated at 12:116-120 (Fig. 106). The sedimentary sequence shows rapid, vertical and lateral variations with diamicton layers, interfingering by gravel, sand and silty sand.

The diamicton layers in the northern part of the excavation consist of thin, wedge-formed and lens-formed beds. The individual beds can generally be subdivided into three vertical zones. The matrix in the lower zone is sandy. The middle zone has a silty, clayey matrix. Both the zones have granules and pebbles dispersed in the matrix. The upper zone has a slight decrease in silt and clay content, compared to the middle zone. The



Fig. 106. Flowed till, gravel lags and fluvial sand at site 12. The sequence is overlayers by laminated silt and clay and basal till. The Laebrink member.

most striking difference is, however, the absence of granule and pebble clasts in the upper zone. The thickness of the individual zones differs between the beds but also within individual beds. The middle part is, however, always significantly thicker than the two other parts.

The exposed portion of the thick diamicton unit has an irregular geometry. It consists of one or possibly two beds. There is, however, no obvious boundary in the middle of the unit except for a 60 cm long sand lens. The diamicton is not completely homogeneous and contains slabs and lenses of sand, especially in the lower part of the unit. The diamicton matrix contains about 50% of silt and clay. The clasts are dispersed in the matrix, and even boulders of 80 cm in diameter are supported by the matrix. A concentration of cobbles occurs in the contact to the sand in the lower part. The sand "in front" of the rounded edge is folded and contains a cone-formed feature with a vertical stratification in the sand.

The interbedded sorted sediments are dominated by two different facies types. One is a well sorted sand or silty sand. The structures in the sand are dominated by planar lamination. Ripple lamination is rare, and the planar laminated sand is often directly overlaid by laminated sandy silt. Load structures are frequent.

The second sediment type is thin layers of gravel. The grain sizes in the layers range from granule to cobble sizes. The clasts are angular to subrounded, and a few clasts with striations were found. The layers are discontinuous, and the lower contacts are conformable and sharp, in a few cases slightly erosional.

Interpretation

The extensive heterogeneous coarse-grained layer shows evidence of several processes. The striated boulders were transported in traction by a glacier. The structures in the layer are not in accordance with a basal deposition, and thus the glacial transport preceded the final shape of the bed. The concentric disintegration of the greenstones and the shattering of the flints were made by physical weathering. The rock pieces have mainly remained in situ, and the reworking or resedimentation of the fragments is restricted. The lenses and sometimes more extensive layers of sand and gravel were produced by selective transport and deposition in current waters. Small shallow rills between the large clasts produced lenses of laminated sand and silt. Larger rills, with increasing speed of current, produced the more extensive layers of sand and gravel. The fluvial activity varied laterally. Boulders and cobbles are relatively more frequent in the parts that show the least flow velocities, i.e. where the silt and sand lenses occur. These parts also contain the diamicton lenses. The coarse material in these parts is considered a residuum of sediments, which originally were transported to the place by a glacier. A slope washing eliminated the fines in the matrix, and only minor lenses of diamicton were preserved.

The parts of the section that contain better sorted gravel were transported to the place by fluvial action. Scattered, buried boulders are, however, regarded as residual clasts.

The thin, laminated sequence on top of the coarse material was deposited during alter-

nating processes of fall-out from suspension and lower flow regime underflow currents in a glacial lake environment. The first sedimentation filled the interstice between the clasts with fine sediments and coated the up-sticking boulders with a clay lamina. Lateral variations in thickness of silt and fine sand cosets were caused by the spatially restricted deposition from the turbidity currents.

The tabular and wedge-formed diamicton layers in the exposure at 12:115-120 were deposited by till flows. The internal variations in texture indicate vertical differences in physical properties of the flows. The lower sandy part has probably got its high sand content by entrainment of sand from the substratum. The absence of clasts in the uppermost part is due to a decrease in bulk density and a reduction of strength. The excess of water from the body has resulted in an upper part of liquefaction and a sinking-down of clasts into the main body.

The thick diamicton body may also be interpreted as a sediment flow. The matrix strength was high, permitting boulders to remain suspended in a rigid plug zone. Sand from the substratum was incorporated in the flow, probably by buckling and overriding. The buckling of sediments in front of the flow can be seen in front of the debris flow body. The sand was folded or completely decomposed by the stresses, caused by the flow. The sand was water-saturated during the deformation, and the pipe-formed vertical structure in the sand was formed as a sand volcano by upward escapement of sand and water in fluidized state. The compression of the sediments in front of the flow increased the shear resistance and made the flow stop. The ridge of the head of the flow was later reworked by new minor flows. The fold structures in the sand indicate a movement from the south-west. The thick flow corresponds to type I flow, described by Lawson (Lawson 1979).

The rapid lateral and vertical variations in sediment texture of the sorted sediments indicate rapid changes in the sedimentary processes both in space and time. The relatively well-sorted sand with discontinuous and dispersed silt laminae was actively sorted and transported by flowing water. The generally fine grained texture indicates low flow velocities below the critical threshold for movement of coarse sand and gravel grains (Hjulström 1939, Sundborg 1967). The plane bed-forms in the sand were deposited in the upper flow regime, which is favoured by very shallow streams or sheet flows (Simons and Richardson 1962, Simons et al. 1965). A transport and deposition of the gravelly sediments in current water is less likely. The large clasts - some of them more than 20 cm in diameter - have very high transport velocities, far beyond the critical threshold for the entrainment velocity of the sand grains in the substratum. Nevertheless, the sand shows no or very slight evidences of erosion. It is more likely that the discontinuous gravel beds are lag deposits, left behind by sediment flows, or that they are a residuum from sediment flows where the matrix has been washed out by current water from rills and sheet flows. This is a common process in low-angled slopes in the ice ablation environment (Lawson 1979).

Environmental interpretations

The unconformity to the lower sediments and the distinct change in grain size composition and petrography composition reflect important changes in the environmental conditions. The

contribution of new material, including striated clasts, is only an indirect evidence of a glacial advance over the area. The facies association at site 12 with diamictons, sorted sand and gravel lag is typical of the sub-aerial resedimentation zone at a glacier margin in a wet, unconsolidated, unstable environment (Lawson 1979). There are no evidences of glacial advance over the area before the deposition and no evidence of buried ice below the sediments when they were formed. The sediments do not show signs of postdepositional dislocations, and the sequence lacks lodgement till and melt-out till facies.

The stony layer, which was traced along the entire cliff, is considered a residual of sediments, similar to the sequence at site 12. The maximum flow extent of different sediment flow types in the terminus region of Matanuska Glacier in Alaska does not exceed 50 m (Lawson 1979). It is not likely that high density flows spread the coarse debris over an area of several hundred m² without the help of an advancing glacier. The advance was probably of a minor order, and the zone of subglacial lodgement deposition did not reach the spot. A short period of climatic amelioration resulted in a temporary retreat. The slope washing, the water-deposited sediments and the till flows show that the excess of meltwater was high. Current water washed out fines in the sediments on the slopes. Small rills coalesced to larger rills in the lower parts of the slopes, where sand and gravel were deposited. It is probable that other periglacial slope processes contributed to the final, almost uniform, thickness of the coarse grained layer. A uniform thickness is at least characteristic of blockfields in periglacial freeze-thaw environments as interpreted by L. Strömqvist (1973) with evidence from northern Norway.

The high water content in the soil is also shown by an active frost cracking and weathering of flints and greenstones. These processes prove to be most intense in clasts, partly immersed in water and with freeze-thaw cycles of "Icelandic" type, i.e. with daily fluctuations around the freezing-point (S. Wiman 1963, G.R. Douglas 1972).

The transition from a subaerial periglacial environment to a glacial lake environment was rapid. A small lake or pond was formed maybe by damming from the glacier. The lake was occasional. The 6 to 8 clay laminae may represent winter sedimentation in the lake but may also be the result of intermittent snow and ice-melting during one single summer or a few seasons. The supply of silt and clay fractions from the ice and the other surrounding environment must have been high. The diamicton debris in the glacier contained about 50% of clay and silt. The surrounding hill slopes consisted mainly of silt and clay from the Glumslöv member. In any case, the lake was of a temporary character and preceded the final Laebrink main advance over the small basin area to the north-west of the Glumslöv hills.

Features of the lower boundary of the fissile diamiction, unit 28.

The lower contact of the fissile diamicton facies of the Laebrink member was studied in several sections. It is everywhere a sharp lithological boundary and comprises informative features, important for the environmental interpretation. Structures, connected to the lower boundary, are therefore treated separately.

Description

The unconformity between the diamicton and the substrata is either angular or parallel. The angular unconformity is the contact to faulted and folded sediments of the Västernäs or Glumslöv members. The distortion of the lower strata can, by directional and stratigraphical indications, be connected to older glacial tectonic events. The contact is generally knife-sharp, and the primary sedimentary structures of the angular strata can be traced to the base of the diamiction (Fig.107). In specific cases, which will be discussed later, a shallow fault system can be seen in the substratum. It seldom affects more than 0.5 m and is closely connected to the lower contact of the fissile diamicton. The parallel unconformity is either a contact to the top strata from the Västernäs member or a contact to lower beds of the Laebrink member. The interface is generally smooth and horizontal with a few exceptions.

One exception is the occurrence of a fossil ice-wedge cast, which was found in the clay pit site 13 (Fig. 108). The wedge was formed



Fig. 107. The knife-sharp lower boundary of the fissile diamicton facies (unit 28) of the Laebrink member.



Fig. 108. A fossil ice-wedge cast, filled with diamicton. From the Strandnäs clay pit, site 13.

from a sand surface and infilled and covered by diamicton. The diamicton in the wedge and the diamicton on top of the wedge are almost identical in petrographical composition (Fig. 49). The clasts in the wedge and the clasts in the lower part of the diamicton on top of the wedge have their long-axes almost vertical. The wedge was 40 cm wide at the top and 2.7 m deep. The diamicton on the top was fissile. There was no sharp boundary between the diamictic infilling and the fissile diamicton on the top.

A slight enrichment of boulders frequently appears at the lower contact. This enrichment exists both when the substratum consists of an older till and when it consists of sand and silt sediments. Two distinct types of boulders and boulder enrichments were recognized.

The first type consists of boulders, buried in the substratum and with a bevelled and striated upper surface, facing the upper till. This type only occurs, when the substratum consists of an older till, and was found in the contact between the Västernäs till and the Laebrink till (E1-facies). The upper bevelled surfaces were striated from the south, while buried parts were striated from a divergent north-east direction. Bevelled and striated boulders of this type also occur in slight concentration zones within the Laebrink till (E1-facies) at site 8.

The second type of boulders was found in the sections at sites 7, 8 and 10. A slight concentration appears independent of the sediment type in the substratum. The boulders are only partly buried, and 30 to 70% stick up into the overlying till. The boulders lack an upper bevelled surface and are generally smooth and striated all way round. Edge-shaped clasts have their snouts directed towards the south, and the northern side is sometimes rough and irregular. The striations and long-axes of the boulders are parallel to clast orientation in the upper till.

The second type of boulders was found in connection with undulating irregularities in the contact between the till and the substratum. These irregularities were studied in two sections, section 7, which strikes in S 85°W, and section 10, which is perpendicular to section 7 and strikes in S 10°E.

The irregularities consist of straight parallel grooves, running almost perpendicularly to the E-W cut in section 7. The grooves incline with a gradient of 8 to 15°, and those, ending within the exposure, terminate abruptly against a boulder or a cobble. The size of the clasts ranges between approximately 15 and 100 cm. As much as 70% of the clasts are buried in the sandy substratum. The sediments beside and below the boulders are faulted, generally by thrust faults with decreasing angles away from the boulder (Fig. 110B).

A 0.2 to 1.0 cm thick layer of laminated sand and silt follows the subsurface of the boulder. The infilling of the grooves consists of diamicton. One example is shown in Fig. 110A. The infilling cannot be separated from the continuous layer of fissile diamicton on top of and outside the groove. A sand streak in the diamicton bends down into the small groove depression. Fabric analyses in the groove and outside the groove show only small differences (Fig. 109). The orientation of clasts shows a unidirectional trend, which is parallel to the orientation of the groove. The clasts in the infilling show a greater variance around the mean, both in a-axes orientation and in the angles of dip.

The section at site 10 is parallel to the grooves. A 9 m long exposure shows deforma-

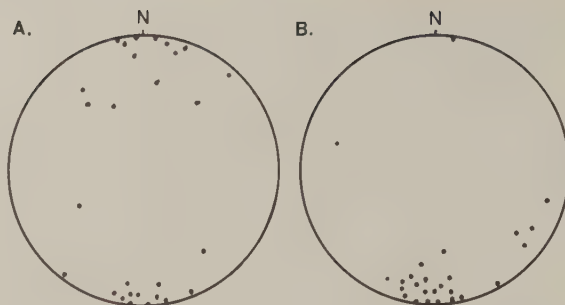


Fig. 109. Fabric analysis in the diamicton infilling of a groove (A) and two meters beside the groove. Fabric analysis from the infilling of a groove (Fig. 109 A), and two meters beside the groove (B).

tion of the substratum (in this case sand from the Västernäs member), a boulder and a wedge-formed groove infilling (Fig. 110C,D,E). The structures are described from the south (to the right in the figure) to the north. The deformation of the substratum starts with a number of low-angle thrust faults. The shear planes strike in E-W, and the angle of the dip increases towards the north. The groove starts at 5.75 m. At this point a series of step-like, high-angle and almost vertical faults are developed in the sand. The steps start from a midpoint fault and rise in both directions. At 3 m two systems of reverse faults with a dip of approximately 45° intersect in a central zone. The uppermost part, 5-10 cm of the sand, is highly sheared along shear-planes, parallel to the subsurface of the groove. The groove is 4 m long and ends abruptly against a boulder. The inclination of the groove is 10°. The boulder is 1.5 m long and 0.8 m high. The edge-formed boulder smooth and striated all way round. Edge-shaped clasts have their snouts directed in front of the boulder has curvilinear thrust faults. The infilling of the groove consists of diamicton with lenses and thin layers or laminae of sand and silt, some of which show slump structures. The infilling is separated from the fissile diamicton on top by a distinct boundary. The small depression on the leeward side of the boulder is also infilled with sand and diamicton. The layers have a concave lens-form, and they start from the upper edge of the boulder.

The two types of boulders represent different processes at the base of a moving glacier.

The buried boulders with bevelled and striated upper surfaces are of a type that is seen in boulder pavements (Dreimanis 1975, p 35, Flint 1971, p 169). The grinding was formed by clasts in the glacier, which moved on top of an autochthonous boulder. The boulder belongs to an older sediment, not necessarily connected to the overriding glacier. The surface grinding and bevelling represent a glacial abrasion phase with a successive lowering of the substratum surface. The boulder enrichment is generally interpreted as the result of a selective removal of small grain sizes, leaving a residue of boulders (Flint 1971, Dreimanis 1975).

The second type of striated boulders represents an abrasion during transport. Velocity differences between a retarding boulder and the surrounding ice, moving faster, result in abrasion of all the sides of the boulder except the lee-side. No bevelling of the top surface is developed.

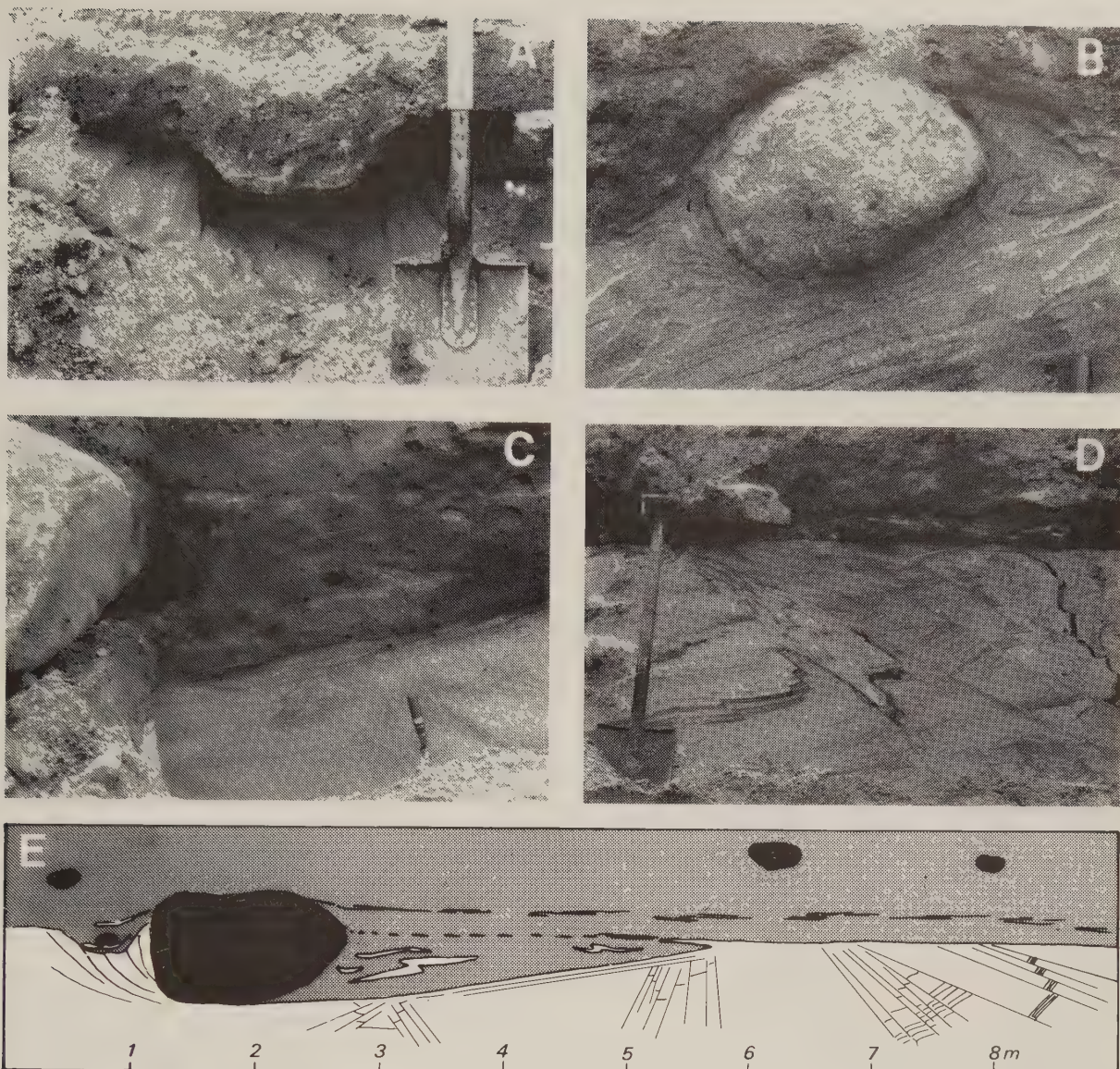


Fig. 110. Grooves and deformation structures in connection to the grooves. A. Cross-section of a groove. Note the down-bending of the sand lens. B. A boulder pressed down in a sand substratum. Note the deformation in the sand. C, D and E. Shear and pressure structures in sand formed during lodging of a boulder, and formation of a groove.

The enrichment of allochthonously striated boulders represents a first deposition from debris in basal transport (Dreimanis 1975, p 35). The grooves with associated allochthonously striated boulders are intimately connected to the lodging phase. Grooves in a substratum of soft, deformable material have been described from the Pleistocene tills in Alberta, Canada, by Westgate (1968). They are interpreted as drag marks, produced by rock fragments that project below the sole of the glacier, as it slides over a deformable substratum (Westgate 1968). Boulton (1975) gives a more theoretical approach. The stress on a boulder, transported in traction, presses it down into the soft, deformable substratum. The progressive consolidation of material in front of the boulder increases the resistance. This, combined with decreasing forces against the upper part of the boulder, as it gets pressed down, makes the boulder finally stop (Boulton 1975).

The deformation of the substratum in the two sections shows no signs of plastic deforma-

tion or intergranular strain deformation. The lamination from primary sedimentary structures in the sand is preserved even in the strongly sheared zone below the groove at site 10. It is most likely that the intergranular movement was inhibited by cementation of ice in the sand. The laminated film around the lower part of the boulders shows that water, produced by pressure melting, did not infiltrate into the sand but was confined to the high pressure area at the boulder contact. The interstitial ice cementation in the sand prevented the infiltration but forced the movement of water around the boulder. The interstitial ice is thought to have been close to or at the pressure melting point.

The deformation of the substratum started when the boulder reached the base of the glacier and was dragged over the substratum. The friction against the substratum retarded the movement. A retardation of the clasts forced the ice to move around the retarded obstacles. The observed boulders were of the size from 20 cm to 1 m. The movement by

enhanced basal creep or pressure melting mechanism is least efficient around obstacles of this size (Weertman 1957, Nye 1974). The retarded clast in Fig. 110E offered great resistance to the glacier flow, and the boulder and the substratum were subjected to a high pressure. The pressure was released by shearing of the sand, and the boulder moved a little bit further before it was retarded again. This time the normal stress from the boulder resulted in a pressing down of the sand (at 5.5 m in Fig. 110E). The boulder ploughed the groove and was finally stopped. The stratified diamicton with sand and silt lenses in the infilling of the groove is interpreted as till flows alternating with sedimentation of sand and silt from current flow. Water was released by pressure melting of glacier ice on the upstream side of the boulder and by melting of interstitial ice in the contact to the sand. Diamicton debris, released from the glacier, and sand and silt, released from the substratum, were flowing into the cavity of the groove. A minor cavity was formed in the upper lee-side part of the boulder. This cavity was also filled with flowing diamicton debris and with sand.

The infilling of the groove cavity as shown in Fig. 110A is of another character. The filling consists of lodgement till, which is a continuous layer along the whole section. The downbending of sand streaks and the slight divergence in clast orientation may indicate a lowering of the till after formation. In this case the groove cavity was infilled by refreezing ice instead of by till flows and sand.

The conclusion is that the glacier substratum contained interstitial ice, the temperature at the base was at or just below the pressure melting point and that refreezing sometimes occurred in the upstream cavity of the grooves. The presence of ice in the substratum also explains the extremely sharp contact to the substratum with preservation of primary sedimentary lamination at the sole of the till (Fig. 107).

The fissile diamicton, unit 28

Description

The fissile diamicton is the most extensive and continuous bed in the area. The bed has a maximum thickness of 8 m, but the average thickness is 3 m. The comparison of texture samples from different parts and different levels shows that the diamicton is very homogeneous (Table 5). The changes in petrography, which were discussed at p. 96, are only reflected by a minor difference in the grain size composition. A lower zone, with texture that evidently diverges from the average composition of the diamicton bed, can be recognized especially. This lower zone ranges in thickness from 0.2 to 1.4 m and consists of crudely laminated diamicton with a major or minor element of silt and sand. The lamination does not have the character of primary sedimentary structures, formed by currents or fall-out from suspension, but is rather a banding or foliation with individual streaks, a fraction of a cm to several centimeters thick. The "layers" are structureless or foliated, and Z-formed folds frequently interrupt the horizontal lamination (Fig. 112A). Individual streaks were sometimes traced for tens of meters.

Striated clasts are frequent. Palaeozoic and Cretaceous limestones are especially sensitive for striation (Fig. 112 B).

The diamicton element increases upwards, and the sediment grades into a more homogeneous fissile facies. At least three different fissure systems were recognized in the diamicton. The three types of fissility are described from cuttings, perpendicular to the ice flow direction.

Type 1 of fissility consists of a subhorizontal fissure system with a maximum vertical space of 25 cm between the fissures. The diamicton cells are elongated, and many of them are up to two meters long. Large clasts are dispersed within the cells, but they never protrude into a fissure plane. The fissure plane is faintly slickensided, and a rough and a smooth direction can be recognized by the finger-tips. The fissures are distinct even in a moist material. A second order fissure system with small, high-angle or almost vertical joints was also recognized. These joints were closed and not noticed in moist parts. The first order fissures are almost horizontal.

Type 2 of fissility shows a more closely spaced fissure system (Fig. 112C). The vertical distance between the fissures is generally between 2 and 5 cm. The general trend in the orientation of the fissures is horizontal in a cutting, perpendicular to the orientation of clasts, and slightly dipping towards the south in a cross-section, parallel to the preferred clast orientation. The fissures converge and form irregular sharp-edged lenses. Axes, drawn from one edge of the lenses to the other, dip between 7 and 15° towards the south. Fabric analyses (Fig. 111) show a well preferred orientation of clasts and a dip, which generally ranges between 0 and 30°.

Type 3 of fissility splits the diamicton into elongated horizontal lenses. The lenses are very regular with a smooth, slightly convex upper surface. The surfaces are slickensided with a close system of striae or small grooves a few millimeters wide (Fig. 112D). Individual grooves were traced for more than 30 cm into the cliff wall. The cutting was

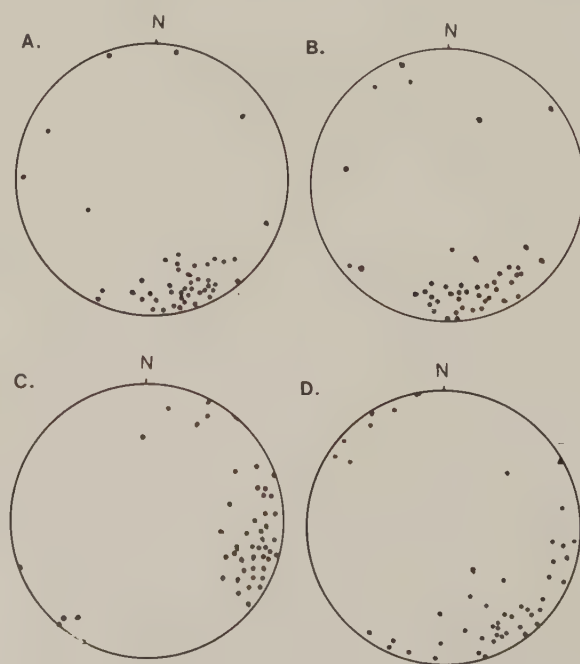


Fig. 111 Clast long-axis orientation in the fissile diamicton. A. From type 1 fissility, B and C from type 2 fissility, D from type 3 fissility.



Fig. 112 The fissile diamicton facies of Laebrink member. A. Z-formed fold in sheared sand-lens. B. Striated clast. C. Type 2 fissility. D. Type 3 fissility. Note the striated horizontal surface in the middle of the picture.

perpendicular to the orientation of the grooves, and the size of the lenses was only measured in one direction. The extremely consolidated sediment and a second order vertical joint system, parallel to the wall, made a cut into the wall, parallel to the elongation, very difficult. The width of the lenses ranges from 9 cm to 80 cm and the height from 1.7 to 16 cm. There is an obvious connection between the width and the height of the lenses, and the ratio between b- and c-axes of the lenses is around 5:1. The b-axes were horizontal. The a-axes were not measured. The slickensided surfaces were almost horizontal or with slight dips both in southern and northern directions. The distribution of clasts is scattered. In some cases, it was obvious that the upper and lower boundaries of a clast coincided with the upper and lower boundaries of a lens-formed cell. The clasts were also centrated to the middle part of the vertical cross-section of the lenses. The fabric analyses show a preferred orientation, parallel to the lineation of the slickensided surfaces. The variation in orientation and dip is greater than in the two other fissure types (Fig. 111).

The vertical and lateral distribution of the different fissure types was not consistently investigated in the sections. The second type fissures are the most frequent. A more thorough investigation was made in the northern cliffs on Ven. Type 1 and type 2 fissility is represented at site 6. Type 2 dominates sections 7 and 8. Type 3 was found in two zones in the lower middle part of the fissile unit at site 8.

Interpretation

Fissility pattern in tills is generally considered the result of subglacial shearing. There is no doubt that the slickensided surfaces in the three different fissility patterns were formed by shearing. The obvious difference in appearance between shear planes in the first two types, compared to the striated and grooved surfaces in the third type, is probably a question of duration and distance of displacement along the planes. The consistency of small grooves in type 3 indicates a considerable movement between the different cells (at least more than 30 cm). Features, similar to the striated and grooved surfaces, were not found in type 1 and type 2 of fissility. The slightly developed slickensiding in these types may then indicate only minor relative movements along the slickensided surfaces.

The geometry of the cells also differs. The streamlined form of the cells from type 3 also indicates a more consistent movement on top of the lenses. The geometry of the cells in type 2 is much more irregular, and the displacement along the shearplanes never adjusted the form of the cells to the movement.

It is likely that the fissure patterns represent two distinct types of basal till formation and deposition. The first two types of fissility were formed in a subglacial position by the overriding glacier, acting on debris below the glacier. The fissure pattern is comparable to the fissures in the Ålabodarna till, where proglacially deposited sediments were obviously sheared by an overriding glacier. The

extremely well preferred orientation indicates that strain was an important part of the deformation during the stress conditions. Intergranular movements are facilitated by the presence of pore-water and high pore-water pressure. The bulk transport or translation of the cells as a whole along shear planes happened when the excess pore-water was drained away and the internal grain interaction was obstructed. The shear stress was then released by brittle fracture and shear movements. The two different fissility types, types 1 and 2, are only gradational differences in intensity of the shear stress during the fracturing, or a question of stress distribution and translation into shear movement in the substratum.

The gradational change in clast orientation in the vertical till sequences indicates that this subglacial process happened more or less continuously and that a till bed was gradual-built up. The consolidation of the substratum increased, and an established preferred orientation was not affected by later changes in ice flow direction.

The successive building up of the bed also indicates a continuous supply of new debris. Sedimentation rates beneath modern glaciers are estimated to vary between 0.5 and 3 cm yr⁻¹ (Nobles and Weertman 1971, Goldtwaith 1975). The two main processes, acting during the formation and the deposition of the till, were thus: basal melt-out and shearing.

The fissility of type 3 is of another type and another genesis. The stream-lined lenses stem better to a mechanism of continuous accretionary lodging, a process which has been described e.g. by Rose (1974), Boulton (1975) and Menzies (1979, 1981). The debris was either transported as aggregates at the ice-bed interface and finally retarded and deposited, or individual particles in basal transport were caught by a clast, already lodged or retarded, until a streamlined aggregate was formed. The ice continued to flow over the top of the cell, and the striae and grooves were formed. The good but less pronounced preferred orientation, compared to the orientation, formed by strain in type 1 and type 2, can also be indicative for the micro-drumlin formation. Diverging fabrics were probably formed on the different flanks of the lenses, and there may also be differences in lee-side and stoss-side accumulation. Such differences within the individual lenses were not investigated but are a theoretical possibility and a possible explanation of the differences in orientation of clasts.

Type 3 of fissility is the result of plastering on processes with alternation of basal accretion and basal sliding. Basal erosion was probably of minor importance. This conclusion is drawn on the consistent relation between height and width of the lenses. The consolidation of the sediments was formed at the same time as the deposition or prior to the deposition. That means that the substratum was well drained during the sedimentation.

6.6 Laebrink member, units 29-32, and Kyrkbacken till: the disintegration of an ice during a marine transgression.

The strata on top of the fissile facies can be divided into two distinct types of facies associations. Simplified models for the two facies transition types are:

Sequence of facies transitions, type 1

Elf (→ E2) → E3D → D

Sequence of facies transitions, type 2

Elf (→ E2) → E3B → C/D → E1,2,3

The strata lie conform to the surface relief and have not been affected by any post-depositional dislocations. This permits a simple comparison of the different facies by their altitude and lateral position. The facies transition of type 1 can be found in the low cliffs on the north and north-west coasts of Ven and in the low cliff north of the Ålabodarna village on the mainland coast. The first facies transition type can be found from a few meters above the present sea level to approximately 10 m a.s.l. The facies transition of type 2 is represented in the high cliffs on the west and south coasts of Ven. The vertical position of the base of the facies sequence is everywhere above 13 m a.s.l. An upward restriction of the distribution of the facies association type was not found in the area, and a minimum value of approximately 40 m a.s.l. represents the height of the highest cliffs on the west and south coasts.

Facies transitions, type 1

Description

The first type of facies transitions was investigated in the Laebrink cliff, site 7 (Fig.29). The stratified diamicton of unit 30 (site unit 7:6) and the silt and clay of unit 31 (site unit 7:7) lie as continuous, subhorizontal layers in the 240 m long section. The units were also traced in the sections at sites 8 and 9 and can be regarded as consistent and laterally extensive layers. The lower boundary of the stratified diamicton at site 7 lies approximately between 4 and 5 m a.s.l.

The geometry of the beds in the stratified diamicton was only studied in two dimensions. The individual diamicton layers range in thickness from 1 cm to 30 cm (Fig.113 B,D). The thick layers (5-30 cm) generally have a restricted lateral extent and were usually traced for less than a few meters. The lower boundary is sometimes erosive, and, in some cases, a thick layer is bounded by a trough-shaped erosive scar. The thin layers are more consistent and show a remarkable uniform thickness over tens of meters. The lower boundaries of the thin layers are non-erosive. The thin layers predominate in the section.

Sampling for grain size analysis was made in the eastern part of the section (Fig. 113A). The excavation contains fissile diamicton facies on top of dislocated sand, stratified diamicton with several thin and one thick

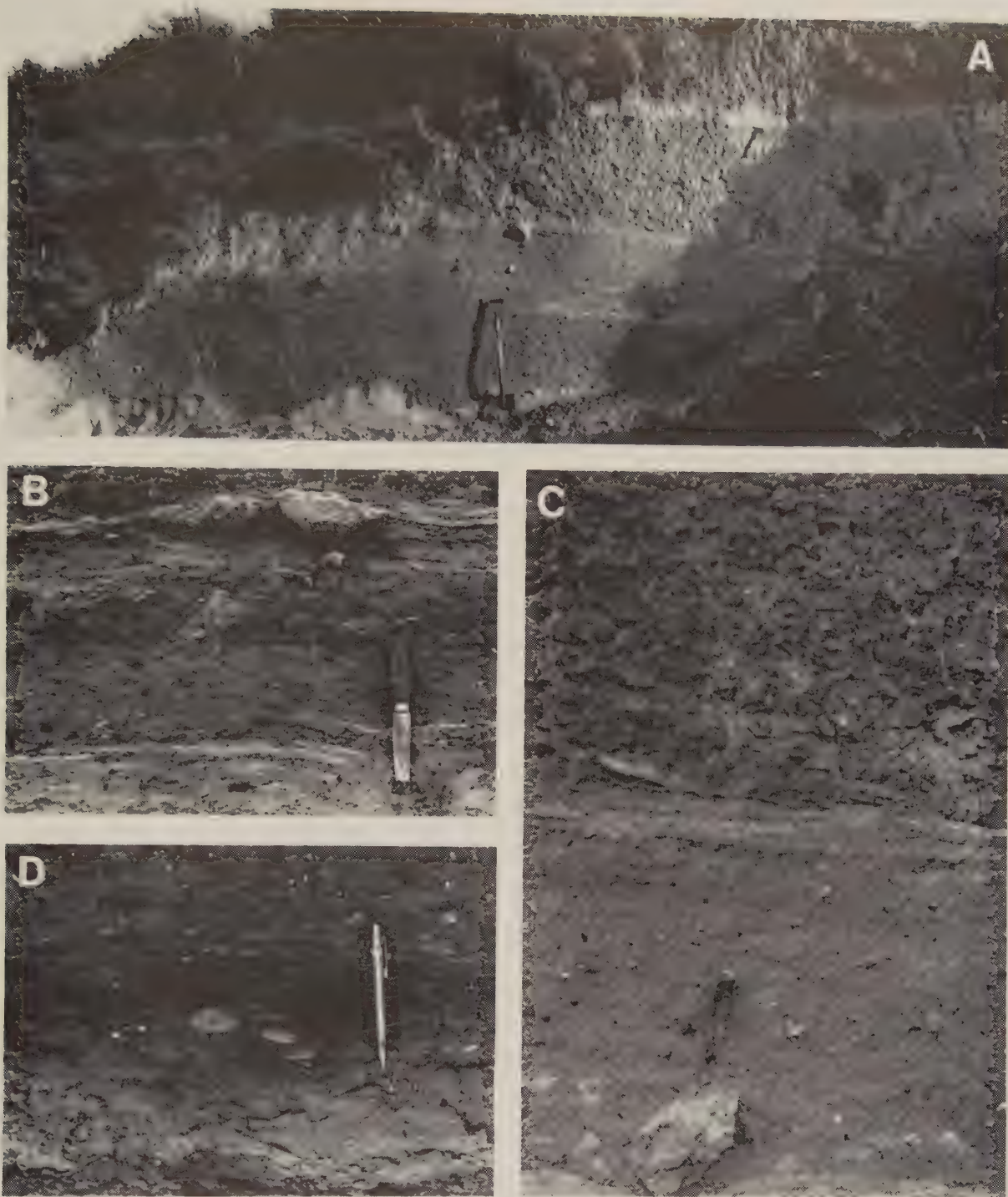


Fig. 113. A. The eastern part of section 7. Strata from the base to the top: dislocated sand, fissile diamicton facies (at the spade), stratified diamicton facies with silt and clay interbeds and clay. B and D. Till flows. C. Till flows overlaid by clay.

layer and with silt and clay interbeds, and finally a massive clay with scattered clast. The grain size parameters of matrix samples are shown in Fig. 114A by the relation between mean values (M_z) and standard deviation (σ_0) and in Fig. 114B by the relation between standard deviation and skewness (α_0). The thick diamicton beds in the stratified facies show grain size parameter values close to the homogeneous fissile facies. The thin diamicton beds have a general decrease of values of all the three parameters, standard deviation, mean size and skewness. A few samples

show an increase in standard deviation and two samples an increase in mean size values.

The clast content also varies between the thick diamicton bed and the thin layers. Clasts, larger than 5 cm in diameter, are restricted to the thick layers and appear randomly within the layers.

Pebbles can also be found in the stratified diamicton as isolated clasts, often with distortion of the bedding in the sediments below the clasts and with a bending of the laminae or layers on top of the clasts.

Laminae of silty sand and clayey silt com-

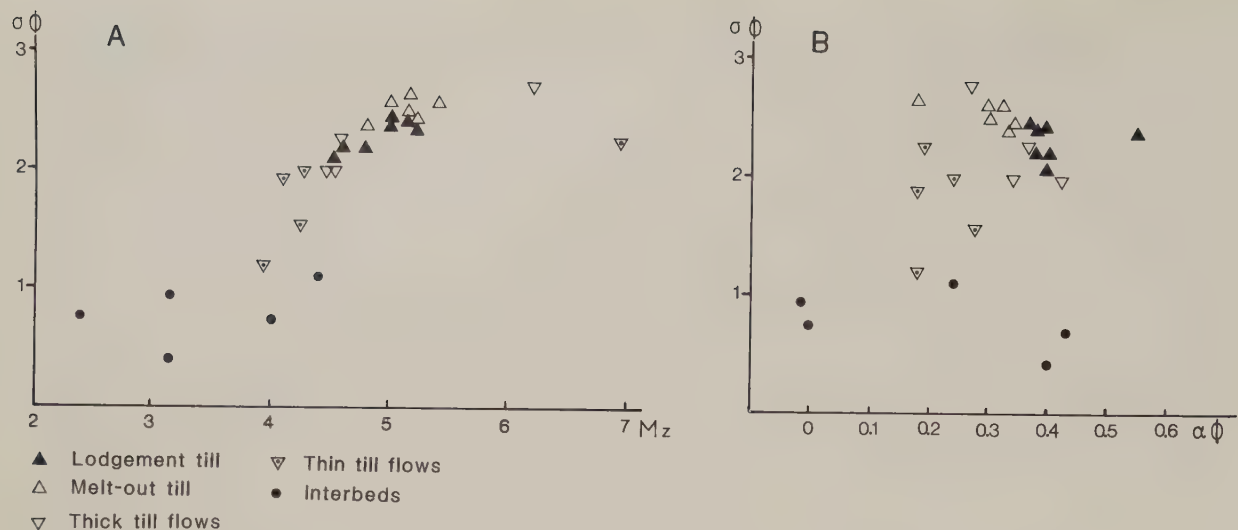


Fig. 114. Grain size parameters of diamicton samples at site 7 and 8. A. The relation between mean values (Mz) and standard deviation ($\sigma\phi$). B. The relation between standard deviation ($\sigma\phi$) and skewness ($\alpha\phi$).

monly occur between the diamicton layers and accentuate the stratification. The laminae are thin, 0.1 to 0.5 cm, but may increase in thickness up to 20 cm. The silty sand is occasionally ripple cross laminated. Gravelly sand and sandy gravel occur as infillings in a few restricted scoured troughs. Consistent clay laminae are found between composite sets of stratified silty sand and diamicton layers. The clay laminae appear in the upper part of the stratified diamicton unit. The thin diamicton layers with increased mean size values, between 6 and 7 ϕ , come from the upper part.

The stratified diamicton is overlaid by clay (Fig. 113C). The clay is massive or crudely stratified. A few consistent white laminae of calcium carbonate were found in the clay. Out-size clasts are scattered in the clay. When dried, the clay splits into cube-formed pieces.

A distinctly stratified or banded diamicton was found on top of the fissile diamicton at 8:270-290. The banded diamicton has the same stratigraphic position as the stratified diamicton and grades laterally at approximately 8:230-250 into the stratified facies. The distinct banding consists of colour variations in different shades of grey, green, red and yellow. The colour banding is not directly connected to any structural features, but there is a distinct subhorizontal fissure system that splits the diamicton into approximately 1 cm thick subhorizontal tabular sheets. The fissures do not converge, and slickenside structures were not found on the fissure planes.

The colour banding can be related to differences in petrographical composition (Fig. 115). The green or greyish green bands contain high ratios of palaeozoic limestone. The red or reddish brown bands contain high ratios of fine-grained red granite and 5-10% of red sandstone. The boundary between layers of different petrographical composition is very distinct. The superimposed bands are not separated by laminae or layers of sorted sediments as is the stratified diamicton in the other parts of the section.

The orientation of clast longaxes shows a very distinct preferred orientation with longaxes in $S10^\circ W$ (Fig. 116). The a/b-planes are horizontal and lie parallel to the tabular banding.

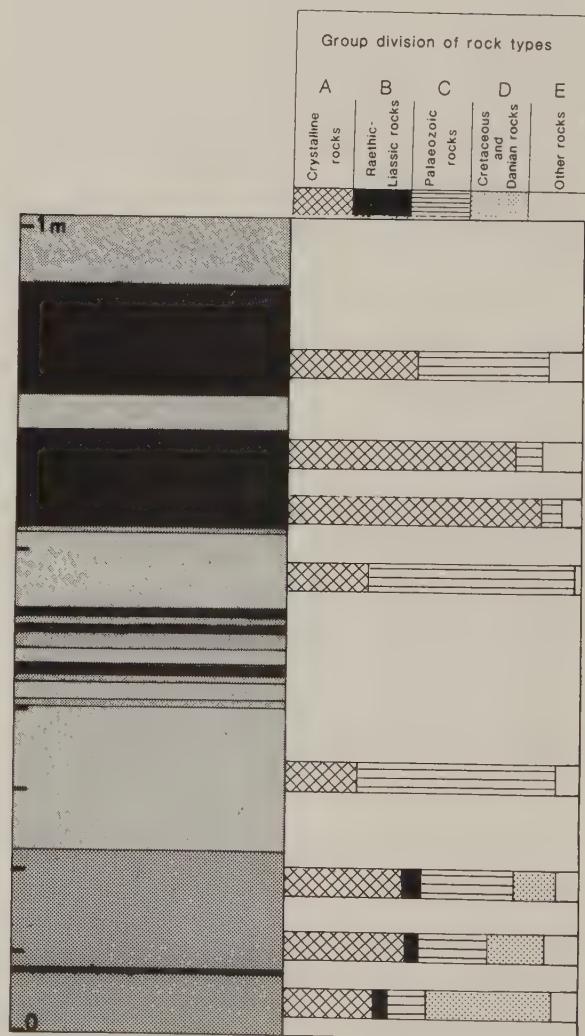


Fig. 115. Petrographical composition of the banded diamicton (melt-out till) at site 8.

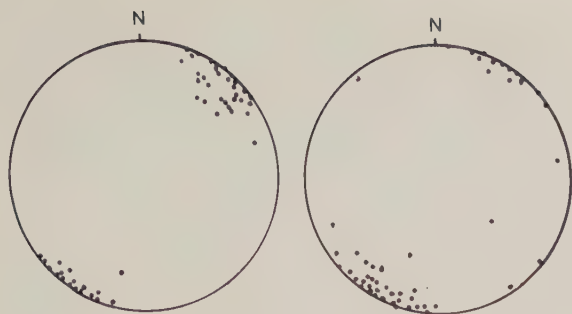


Fig. 116. Clast long-axis orientation in the melt-out till at site 8.

Interpretation

The stratified diamicton facies at site 2 is interpreted as till flows alternated by lag deposits and fluvial deposits. The debris of the till flows was most likely released by melt-out from the glacier ice. The source material is thus supposed to have been of approximately the same grain size composition as the basal till below.

The beds in the stratified diamicton facies are generally coarse grained and the sediments are impoverished in silt and clay size grades. The associating sorted sediments are mainly high energy sediments. The planar laminated sand was, for example, deposited in the upper flow regime, which prerequisites a high flow velocity or a low depth for formation (Simons and Richardson 1961). The general loss of fine size grades is a pervading characteristic of unit 29 and means principally important differences in the distribution of facies below and above approximately 10–13 m a.s.l. Persistent silt and clay sediments above this level are absent. This indicates that the level corresponds to a base level and that the stratified diamicton facies on top of the fissile diamicton in the western and southern cliffs represent subaerial conditions. The consistent clay laminae in the stratified diamicton units at site 7 indicate a subaqueous environment with clay particles falling out from suspension. The ripple cross-laminated silt and silty sand was formed by low velocity currents. The subaqueous environment, interpreted from the clay laminae, means that the currents were underflow turbidity currents. The thin beds also show that the currents were short-lived.

The deposition of silt and clay alternated with a deposition of diamicton layers. The diamicton layers are interpreted as till flows. Textural variations within and between individual diamicton layers are characteristic of the resedimentation debris flow process at a glacier margin (Lawson 1979). The predominating thin flows with increased sorting, compared to the lodgement till below, indicate a high water content in the debris/water slurry matrix. The slurries were spread out on the basin floor. The thin silty partings between the individual beds were probably deposited from low density turbidity currents or from the suspended cloud formed in connection to the debris flows.

The textural change of the thick flows is more restricted. The similarity to the lodgement till is striking, and the loss of clasts by sedimentation in the flow or of fine grain sizes by intermixture with overlaying water was small. The matrix strength of the thick flows was higher than in the thin, more

fluid flows.

The dispersed clasts, which could not be connected to the debris flows but were situated between the layers sometimes conformably bent over the clast, were deposited as single clasts. They can only be explained by drop from above and are evidences of iceberg transport.

The stratified diamicton is overlaid by clay. The transition from a debris flow sedimentation to clay sedimentation was rapid. The clay is almost massive and evidence of underflow currents could not be distinguished. The scattered clasts in the clay are probably of the same origin as the single clasts in the lower unit, which means that the clasts were settled down in the water after melting out from floating ice.

The most striking characteristic of the banded diamicton in the eastern part of section 8 is the distinct banding which reflects layers of quite different petrographical composition. Such a banding prerequisites very specific erosional, transport and depositional processes. The rock types, found in the upper part of the diamicton, outcrop in the Baltic Sea area, and the transport distance is approximately 300 to 500 km. The mixing between the different rock types is very small. A transport without mixing is possible in the thick basal zone of a glacier or ice sheet with a net basal freezing. The transport of debris in distinct debris bands is found in sub-polar or polar glaciers (Boulton 1970, Shaw 1977, Lawson 1979). The debris bands are formed by freezing-on of water to the base of the glacier, and the debris is entrained during the freezing. This prerequisites transitions from warm-based conditions to cold-based conditions where the water flows from unfrozen parts into the freezing zone. Repeated changes of thermal regime at the base of the ice during further movement over a heterogeneous bedrock area are reflected by vertical changes in the composition of the debris in the ice. The most distant bedrock, which was first entrained at the base, is found in the upper part of the banded ice.

The banding of the ice was preserved during the transport. That means that a net basal melting did not occur during the long transport. During the deposition, several of the properties, derived from the entrainment and the transport, were unaltered. The distinct banding was preserved, a distinct orientation of clast longaxes was preserved, and the textural composition was also preserved. This is possible only by a slow basal meltout process from a stagnant ice. There are no evidences of water escapement or flowing of water in cavities within or underneath the ice. The preservation of the englacial properties was favoured during the melt-out process by the fact that the stagnated ice was submerged. The pressure on the wet sediments after melting out from the ice was probably low due to the reduced weight of the submerged ice body.

The melt-out till is not overlaid by till flows, which is common in subaerial environments. The melt-out till grades, however, laterally into till flows, and the upper boundary of the two till types forms a continuous, horizontal boundary to the massive clay.

Facies transitions, type 2

Description

The strata at site 2 are representative of the second type of facies transitions. The sequence in Fig. 117 comprises the uppermost part of unit 2:6, unit 2:7, unit 2:8 and the lower part of unit 2:9, which corresponds to units 28, 29, 31 and 32 in the correlation chart, Fig. 62.

The lowermost unit is of the fissile diamicton facies, and the upper boundary is slightly scoured. The stratified diamicton unit consists of distinct layers of sorted sand and gravel, poorly sorted sandy gravel and extremely poorly sorted diamicton. The diamicton layers are very heterogeneous in grain size composition, and the mean grain size values of matrix samples from different layers range from 2 ϕ to 4.5 ϕ . The individual diamicton beds were traced laterally and were found to have a very restricted distribution, most of them being less than approximately 4 m. Those, ending within the excavation, had wedge-formed or thickened snouts.

The complex unit changes laterally to a layer of poorly sorted or moderately sorted sandy gravel with scattered boulders. Such lateral transitions are common in the sections on the west and south coast of Ven, and the thin layer of gravel has, as a matter of fact, a larger lateral distribution than the more complex sequences of stratified diamicton. The sand at 1.4 m in Fig. 117 has planar lamination with a 5 cm thick unit of cross-bedded sand with a low angle dip of the foresets.

The stratified diamicton is upwards succeeded by ripple cross-laminated sand, laminated silt and two thick deformed beds, comprising zone A and zone C (the lower one) and zones A, B, C and D (the upper one) of deformed and dissolved strata. (For description of the zones, see p. 65).

The deformed beds are overlaid by clay. The clay contains dispersed sets of cross-laminated sand, but the clay and the sand sets do not form distinct couplets. The top of the sequence consists of diamicton. The diamicton is homogeneous except in the lowermost part. The boundary to the clay below consists of a zone of sheared clay with low-angle shear planes and a zone of brecciated clay. The lower part of the diamicton also contains numerous clay intraclasts, and the average clay content in the matrix is very high. The homogeneous, upper part of the diamicton has a distinct fissure system with 2 to 5 cm vertically spaced horizontal fissures (Fig. 118 D, E). The fissures converge.

The uppermost diamicton unit is the Kyrkbacken till. The unit, as described at p. 54, can be traced as a continuous 3 to 5 m thick bed in the uppermost part of the cliffs on the south and west coasts of Ven. The unit shows lateral variation in structure and texture and was also studied in a fresh landslide scar at site 4 (4:100-150). Fig. 117 shows part of the wall. The unit can be subdivided into a lower, distinctly stratified part and an upper part with a diffuse stratification. The distinct stratification in the lower part consists of diamicton layers, interlayered by clay, silt and sand. The stratification is accentuated by white layers of chalky clay. The diamicton layers are approximately from 1 cm to 10 cm thick. Thicker beds are rare. The layers intertongue with the sorted sediments and generally end as a wedge. (Fig. 118 B,C). Small scale slump structures are frequent in the distinctly stratified part.

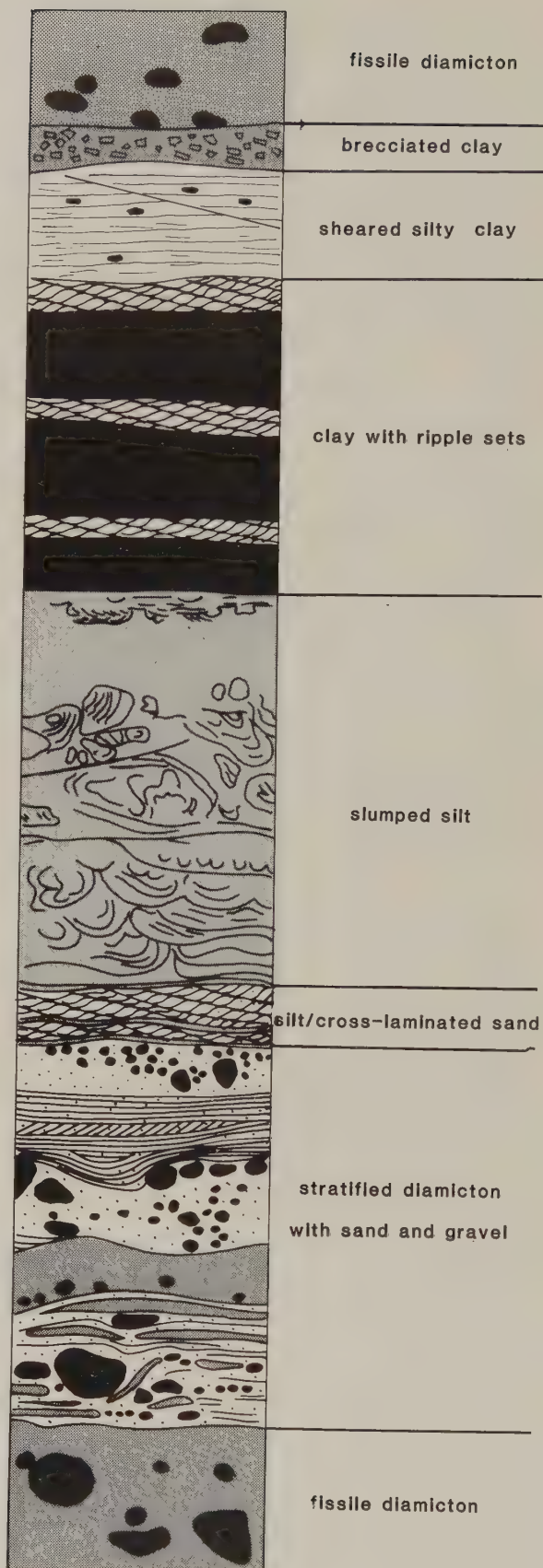


Fig. 117. Log of Units 2:6, 2:7, 2:8 and 2:9 at site 2.

The upper part of the diamicton has a cruder stratification. The individual layers are not consistent and are often interrupted

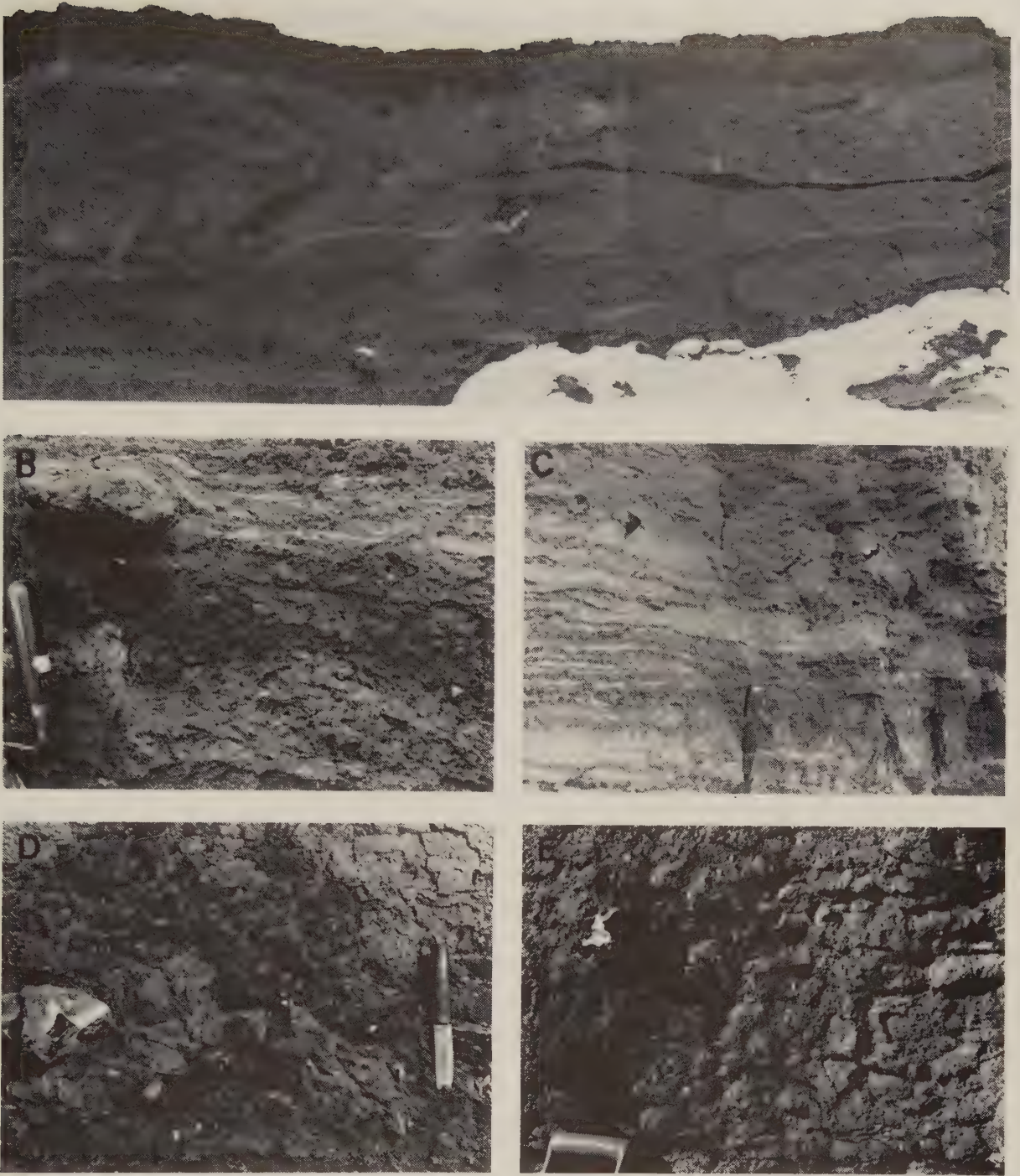


Fig. 118. Kyrkbacken till. A. The crudely stratified and distinctly stratified diamicton at site 4. B and C. A close up of stratified diamicton. D and E. The fissile diamicton facies.

and folded. The distinct layers in the lower part can be traced into the crudely stratified upper part (see in the left part of Fig. A). The layers are overturned, and the distinct stratification is dissolved into a crude stratification. The fold-axis strikes in S 50–60°E.

Interpretation

The sand on top of poorly sorted sediments was deposited from currents of lower

flow regime. The ripple sets are interlayered by laminated silt and clay. This prerequisites standing water or very low flow velocities. The ratio of silt and clay grain sizes increases upwards, and the uppermost of the deformed layers is dominated by these grain sizes.

This transition indicates a change in the environmental conditions to a probable position below the base-level. The subaqueous sediments slumped on the floor, and the upper part was dissolved and fluidized. Cross-laminated sand, grading upwards into parallel-

laminated silt, was deposited, probably from a turbidity current on top of the slumped masses, and the turbidite partly sank into the fluidized sediments.

The clay was deposited by fall-out from suspension. The slow sedimentation was interrupted by currents, building up ripple sets on the basin floor. The currents were most likely density underflow currents.

The sedimentation in the still-water basin was suddenly changed. This is shown by the supply of diamicton debris. The mixture of grain sizes from clay to boulders indicates a transport by glacier ice. The shearing of the clayey sediments below the diamicton shows low normal forces, relative to the horizontal forces, and the deformation of the clayey sediments was most likely made by the ice, which moved into the place.

The section at site 4 shows a very distinct stratification and an alternation between till flow sedimentation, forming the diamicton layers, fall-out from suspension, forming the clay and silt laminae, and layers and currents, forming the stratified sandy sediments. The crudely stratified diamicton with slabs and lenses of sand and white chalk was originally probably of the same type as the stratified diamicton. This can be seen in the overturned strata where a distinct stratification grades into a crude stratification or even a homogeneous diamicton. The deformation of the sediments can thus be connected to a stratigraphic level on top of the deformed and partly dissolved stratified diamicton sediments.

The overturning of the sediments shows an uphill movement. This was not made by gravity, and could only have been made by ice.

6.6.2 Environmental interpretation, units 24-32

The environment during the Laebrink ice advance was a permafrost area. The ice-wedge at site 13 on the mainland was active, and the sand at site 10 was frozen.

The ground remained frozen underneath the glacier front, and during the first basal deposition there still remained at least patches of frozen ground. The lodged boulders were interpreted as being left behind by the moving glacier on a substratum of frozen sand, and the ice-wedge did not melt until the deposition of lodgement till had started.

The deposition of lodgement till indicates a basal temperature at the pressure melting point, and it is thus evident that interstitial ice existed in patches beneath the warm-based glacier.

The first ice movement, which has been registered from the area, was from S. This is shown by the trend of grooves and by the orientation of clast longaxes in the lowermost part of the lodgement till.

The lodgement till shows an upward change in the orientation of clasts to a more easterly direction. The change is gradual, and directions, all the way from S to E, are present (Fig. 61). The gradual change shows that the base probably remained at the pressure melting point, or at least that the base from time to time was warm during the gradual change of the ice movement. Nothing can be said about how long time the warm-based conditions prevailed. It is not even clear if it was in the beginning or in the final stage of the glaciation. Evidence of basal temperature of the ice sheet can also be interpreted from the deglaciation sequences.

The deglaciation sediments, both the melt-

out tills and the flowed tills often show a petrographical composition that markedly differs from the lodgement till below. The rock types show a long transport distance, and the absence of mixing between the different rocks indicates that the transport happened in one step from the source to the investigated area. This is of importance for the interpretation of the glacier regime and shows that the transport of debris from the Baltic Sea area to the Ven-Glumslöv area happened in an ice sheet with very little basal melting. The debris actually did not melt out from the glacier before the final stagnation of the ice.

The melt-out till shows that the deglaciation was by stagnation of the ice. This is often seen in marginal zones of compressive flow. The deglaciation sequence of the Laebrink member does not, however, show the characteristic features of glaciectonics, due to compressive ice stacking. Melt-out sequences were found not only in the section, earlier described, but also at site 1 and site 4. In all cases the characteristic banding lies horizontal to the substratum.

The two distinct types of facies transitions, found in the deglaciation sequences of the Laebrink member, are the result of sedimentation in two different environments, a subaqueous environment and a subaerial environment. The base-level had a position somewhere between 10 m a.s.l. and 13 m a.s.l.

The silt and clay of unit 31 is an integral part of the two facies transition sequences. In the low-lying, northern part of the area it overlies subaqueous deglaciation sediments, and in the elevated southern and western parts it rests on subaerial sediments. At site 3 the deglaciation sediments are two-parted, units 3a:2 and 3a:3. Stratified diamicton with silt and clay interbeds lies on top of stratified diamicton with sand and gravel interbeds. The two diamicton facies at site 3 are situated at approximately 18-20 m a.s.l.

The transition into a subaqueous sedimentation at levels above 10-13 m a.s.l. shows an elevation of the base level. The transgression affected the whole investigated area, and Ven was completely covered by water. The maximum level of the transgression cannot be evaluated from the investigated sequences. The clay sediments do not show a distinct lamination, indicating a glacial lake environment, and the decay of the ice implied that the passage to the sea water in the Kattegatt was free. The transgression was thus a rise of the sea level, relative to the land.

The increase of the water depth in the area may have resulted in a conversion of stagnant ice masses to floating icebergs. The distinct and rapid transition from till flow sedimentation to the sedimentation of clay with dispersed ice-dropped clasts may indicate this.

The sedimentary sequences on the west and south coasts contain an additional unit, the Kyrkbacken till. The interpretation of the structures in the diamicton is that the massive diamicton at site 2 was sheared by ice and that the thick sequence of subaqueous till flows was overturned and folded also by ice. The transport of diamicton debris and the shearing thus indicate that the southern and western parts were invaded by ice. The massive diamicton and the overturned till flow sequence belong to the same unit, traced as a continuous unit along the cliffs. It is thus likely that the massive diamicton is of the same origin as the stratified diamicton, and the structural difference is due

to different degree of internal strain deformation.

There are two possibilities to interpret the genesis of the ice. The first is that a glacier invaded the area from the south. The glacier never reached the low-lying area in the north. The second possibility is that the debris was transported by icebergs, and that icebergs, passing the submerged island, grounded on the elevated part of the island. The icebergs unloaded their debris, and the shearing was then made when the icebergs were

dragging on the shoal. The water depth in the northern part permitted the icebergs to drift away, and only occasional clasts were dropped down. The interpretation is analogous to the interpretation of the Lund till in the southwest of Skåne (Lagerlund 1980, Berglund and Lagerlund 1981).

The Kyrkbacken till is generally overlaid by Ögnäs sand or constitutes the surface soil in the cliffs. The sand is a humic wind-blown sand, mainly deposited during historical time.

7 Regional implications

The main purpose of this thesis was to establish a lithostratigraphy of the glacial deposits in the Ven-Glumslöv area and to reconstruct the palaeoenvironmental conditions during the deposition of the distinguished lithostratigraphic units. This was dealt with in the preceding chapters. In this chapter problems and principles of stratigraphic correlations to adjacent areas will be briefly discussed. The palaeoenvironmental conditions of regional implication, which were described in the stratigraphic synthesis, will be emphasized.

Methods for absolute dating of glacial sequences in terrestrial environments are by now not existing, except for the controversial thermoluminescence dating method. The severe climatic conditions at a glacier margin reduce the biological activity, and the environment at the glacier and underneath the glacier can be considered an environment, dominated by physical processes. The absolute dating by using the radiocarbon method is thus dependent on organic sediments or sediments with organic detritus, deposited prior to and after the glacial sequence. This gives only maximum and minimum ages for the glacial sediments.

A correlation of the palaeoenvironments in the Ven-Glumslöv area to the North European chronology and the Quaternary stratigraphy of Norden (Mangerud et al. 1974) must thus be based on the relative position of the sediments, compared to radio carbon dated interstadial or interglacial sediments.

Such a correlation can be made from the sediments of the Uranienborg member. The thick sand sequence is correlated to the extensive sand unit, which is traced along almost the entire Alnarp Valley (see Fig 2). The sand has been interpreted as a fluvial sand (Holst 1911, K Nilsson 1971, 1973), and the river is generally called the Alnarp River (Holst 1911). Attempts to give the sediments an absolute age was made by borings in the south-east part of the valley (K. Nilsson 1971, Miller 1971, 1977). Fine sand sediments were connected to two different chronostratigraphic levels, sediments of a probable Holsteinian age in the Hyby core and sediments of Weichselian age in the Gärdslov core (Miller 1977). Five radio carbon datings from Gärdslov gave finite ages from 21.305-3000 to 32.880-1770. The dating of the Gärdslov Beds (formal lithostratigraphic classification after Lagerlund 1980) indicates fluvial environment in the

Alnarp Valley during late Middle Weichselian and early Late Weichselian. The defined units of the Gärdslov Beds consist of a thick sand bed and an upper silty clayey bed. The same transision from a thick sand bed to a clayey bed was found in the Uranienborg member. The correlation of the Uranieborg member to the Gärdslov Beds is thus based on the similarities in the physiographical and stratigraphical position and the lithological characters of the sediments. Consequently, the sediments in the cliff most probably belong to the Late Weichselian ice sheet advance.

This provides possibilities for correlations to the Weichselian deposits in other parts of Skåne and in Denmark. This has also been made earlier by several authors (see chapter 3), and a traditional correlation of the till stratigraphy of the Ven-Glumslöv area to the ice advances, recognized in Denmark, does not afford any problems (Table 7). The correlation is based on lithostratigraphic units, which are considered time transgressive. A palaeoenvironmental and palaeogeographical reconstruction demands, however, a correlation of time equivalent units. The environmental conditions at the ice border zone in Middle Jutland during the Main advance had probably very little in common with the environmental conditions during the deposition of the Västernäs member in the Ven-Glumslöv area, and, the main objections, the deposits in the two areas are not of the same age. Therefore, correlations, like the one in Table 7, are not adequate in palaeoenvironmental reconstructions.

In order to understand the dynamic of the ice sheet and to understand the interplay between the physical environment and the climatic conditions it is necessary to find out the regional changes in the environment, connected to an absolute time scale. A comparison can be made to the biotic zones and the migration of biotic zones in palaeoecological reconstructions. The intention would be that the glacial environments can be added to the biotic zones, recognized in areas at a distance from the ice sheet, and that the migration of glacial environmental zones can be evaluated and parallel connected to the migration of the biotic zones.

The purpose of the reconstruction of the glacial environment is thus to find out e.g. what the environmental conditions in the Ven-Glumslöv area were during the maximum extent of the ice sheet in Middle Jutland. Such a

Table. 7. A traditional till stratigraphic correlation between Denmark and the Ven-Glumslöv area. The correlation is based on timetransgressive units and has no relevance in a palaeoenvironmental reconstruction. (The Danish ice advances are after Houmark-Nielsen. 1981).

Denmark		Ven-Glumslöv	
Ice advance	Directional indication	Local stratigraphy	Directional indication
The Norwegian advance	N	Ålabodarna till member	N
The Main Danish advance	NE	Västernäs till member	NE
The East Jutland advance	E		
The Baelthav advance	S-SE	Laebrink till member	S-E

correlation cannot be made on the basis of the un-dated lithostratigraphic units, but it can be made on the basis of an environmental interpretation and an evaluation of conditions with regional implications. The literature is poor in genetic and environmental interpretations of glacial sediments with a well defined chronostratigraphic position, and the interpretation of the regional environmental implications are therefore essential for palaeoenvironmental reconstructions of large areas in the future.

From the discussion above follows that the regional implication of the environmental conditions during deposition of un-dated glacial sediments is the most important possibility to make time-equivalent correlations to adjacent areas. Therefore, some of the most striking regional environmental conditions of the different members are discussed.

Uranienborg member

The correlation to the Gärdslov Beds indicates fluvial conditions in the Alnarp Valley during late Middle Weichselian and early Late Weichselian. The regional implication has been discussed by Lagerlund (1980), and he concluded that the fluvial environment prerequisites a sea level at or below the Valley floor.

Ålabodarna till member

The transition from fluvial sediments to glacial lake sediments, as interpreted from the Ålabodarna member, means a change to a glacial lake and a rise of the lake level from -60 m to approximately equal to the present sea level. The Alnarp River was drained to the Atlantic through the Kattegatt Sea, and a lake phase in the Öresund area implies a damming by ice in the pass-point area or a subsidence of the area, relative to the pass-point area.

The indication of glacial deposition in the Ålabodarna member stems well with an assumption that a glacier reached the low-lying pass-point area at Limfjorden. This is also postulated by Lagerlund (1980) in his interpretation of changes from a fluvial environment to a lacustrine environment in the Alnarp Valley during the transition from sandy to clayey sediments in the Gärdslov Beds. The damming must have affected the low-lying parts of the catchment area of the Alnarp River, and evidence of a lake level transgression can thus be excepted in large areas, and can be regarded as a synchronous event in the whole region.

The debris was interpreted as being transported and mixed along the sole of a warm-based glacier. The temperature gradient from the base to the glacier surface cannot be evaluated, and the glacier could have been as well below as at the pressure melting point at higher levels. The heat flux from the substratum was, however, large enough to compensate an eventual negative gradient in the glacier and keep the base wet.

The thick strata of flow sediments, deposited during the deglaciation of the Ålabodarna ice, indicate that the glacier was actively moving forwards, continuously bringing new debris to the glacier front. The glacier was active, at least as long as it remained in the area.

Glumslöv member

The glacial sequence of the Ålabodarna member is succeeded by glacial lake sediments and, later, by sand-dominated braided stream deposits.

Several of the environmental conditions, discussed in chapter 6.2.2, are of a regional implication.

The changes in the lake level must reflect changes within the entire basin, which probably had quite a large extent. A try to reconstruct the extent of the lake was made by Lagerlund (1980). In the Nyhamnsläge glacial lake, as he calls the lake, low-lying areas along the south-west Swedish coast were submerged, and also the coastal parts of Jutland and Sjaelland were inundated.

The reason for the changes can be discussed out from the previous mentioned reason for the origin of the lake. That means that the drop in the lake level might be connected to the withdrawal of the ice from the pass-point area and that the transgression means a new readvance. This is speculations, and other possibilities can also be considered. One possibility is that the basin was gradually sinking, relative to the pass-point area. The repeated slides at the bottom of the lake may indicate disturbances in the basin itself.

A subsidence of the area might be expected during the advance of an ice sheet. A proglacial depression (Walcott 1970) would also result in a transgression if the pass-point area was not equally affected by the subsidence. Both the hypotheses can be supported by the further development in the area. The transgression was suddenly interrupted, and the sedimentation ceased. A final withdrawal of the glacier from the pass-point area resulted in a draining of the glacial lake.

The hypothesis of a proglacial depression can be supported by the fact that the sedimentation of the Glumslöv member is succeeded by a glaciotectionic deformation. A glacier came across the glacial lake and delta environment and dislocated the sediments. Further hypotheses are conceivable but will not be discussed.

The environmental implication of the glaciotectionic deformation was discussed in chapter 6.3.4. The different possibilities in interpreting the physical conditions of the sediments during the deformation have a great importance for the interpretation of regional implications. If the sediments were permafrozen during the dislocation, it means that a severe cold prevailed. The deformation does not necessarily prerequisite permafrozen ground. The glaciotectionics, however, implies a glacial advance, which in itself means a climatic deterioration.

Västernäs till member

The Västernäs member lies on top of the dislocated strata. The unconformity between the dislocation and the Västernäs member was discussed in chapter 6.4.2 and offers possibilities on speculations of glacier temperature. An unconformity is an evidence of erosion or non-deposition. Erosion may indicate, but does not necessarily mean, warm-based glacial conditions. Erosion may also occur underneath a cold-based glacier or a base of transitional character by a freezing-on mechanism. A non-deposition may also indicate cold-based conditions with a base frozen to the substratum.

The strata from the Västernäs member represent deglaciation conditions. Lodgement till was deposited beneath the glacier, and debris from stagnating ice melted out. This evidently means a climatic amelioration which must be of a regional implication. The ice-front oscillated during the retreat, indicating an active ice during the deglaciation.

Laebrink till member

The lower contact of the lodgement till of the Laebrink member contains indications of frozen ground prior and temperaneous to the very first lodgement till deposition (chapter 6.5.1 and 6.6.2). This difference in the environment, from a melting ice during the Västernäs phase to the formation of permafrozen ground, implies a new climatic deterioration. The sediments below the lodgement till at sites 11 and 12 (chapter 6.5.1) might be an indication of a humid, cold climate during the ice advance. The glacier base was, at least part of the time, at the pressure melting point. This is indicated by lodgement till. The petrographical composition of the deglaciation sequences does, however, imply that a net basal melting did not occur during the transport of debris from the source in the Baltic Sea area to the investigated area.

The environmental conditions during the deglaciation of the Laebrink glacier were very specific. The sea level can be determined to 10–13 m a.s.l. (corrections for isostatic response are not made), and the northern part of the area submerged during the deglaciation. The deglaciation in the elevated southern and western parts of Ven started as a subaerial deglaciation, but a transgression gradually inundated the area. The total amount of the transgression cannot be evaluated from the investigated sections. The highest parts of the cliffs are about 40–45 m high, and the transgression was recognized up to this level. Geological mapping of the northern part of the Glumslöv hills (Adrielson et al. 1981) shows, however, that the marine limit is situated at approximately 57 m a.s.l. That means a shore displacement in the area of about 45 m.

This transgression must have a regional signification, and it can be correlated to a transgression with a similar amount in south-west Skåne, north-west Skåne and the southern

part of Halland (Lagerlund 1980). A transgression of approximately 25 m was interpreted by radio carbon dated molluscs from northern Jutland (Petersen 1983). The time for the transgression was approximately between 14.000 and 13.400 yrs BP.

The decay of the ice was by stagnation (see chapter 6.6.2). Melt-out till in combination with flowed till shows that stagnant ice masses were surrounded by ground or sea floor where till flows were deposited. The transgression probably caused some of the stagnated ice to become icebergs. The drift of icebergs probably had effect on areas outside the investigated area. The very specific petrographical composition of the glacial debris in the ice may make it possible to trace the path of the iceberg drift further along the coast.

Kyrkbacken till

The final stage of the glacial environmental history of the area was the deposition of the Kyrkbacken till. These icedrift sediments were discussed in chapters 6.6.1 and 6.6.2. The two possible explanations of the sediments, iceberg sediments (the interpretation introduced by Lagerlund 1980 for Lund Till) or glacier ice sediments have implications for the interpretation of the end of the Weichselian glacial history in the region.

The reconstruction of palaeoenvironments is a fascinating field of study.

An evaluation of the regional validity for the environmental conditions in a small area is necessary and important for an expansion of the environment to a region. The interpretations and the regional implication of the environmental history for the Ven-Glumslöv area contain many alternative conclusions, and it is necessary to deepen and broaden the lithostratigraphical and palaeoenvironmental studies to get closer to a realistic picture of the dynamic glacial environment during the Late Weichselian.

Acknowledgements - This investigation was carried out at the Department of Quaternary Geology, University of Lund, Sweden.

First of all I would like to thank Björn Berglund, my supervisor, for his patience with a self-willed student. The advantages of having his support and confidence widely outweighed the disadvantages of his restricted knowledge of proglacial resedimentation processes and braided stream patterns.

Special thanks are conveyed to Erik Lagerlund, who critically read my manuscript, and whose professional judgement I completely trust.

My most sincere thanks go to two persons, my father-in-law, Erik Bergenrud, and my friend and colleague, Mervi Hjelmroos, for their outstanding help during the last two months. Erik Bergenrud corrected my English and typed part of the manuscript, and Mervi typed the other part of the manuscript. Their practical help has been a prerequisite for the carrying through of this thesis, and their anxious inquiries about more job to be done

have been a splendid motive power to continue the work.

I am also grateful to my colleague, Max Åmark, who coincidentally wrote his thesis in the room next-doors. His enthusiastic talks about sunny days on a surfing board kept me in high spirits during a hard time. His hints and directions concerning the choice of screen and inkpen were professional. Max compiled the reference list, and he and his wife, Catherine Dupuy, also helped me with some of the drawings.

I would also like to thank Svante Björk for reading part of the manuscript, Eva Adrielsson, Lena Barnekow and Inger Lander for reading the diagrams, Roger Lundholm for doing an excellent job with the photographs, and Helena Bergenrud and Tony Åström for their help in the trying job of mounting the pages.

Finally my love goes to Per, my husband, and Martin, 4 years old, for their sacrifice, patience and perseverance.

It will be a lovely summer this year!

References

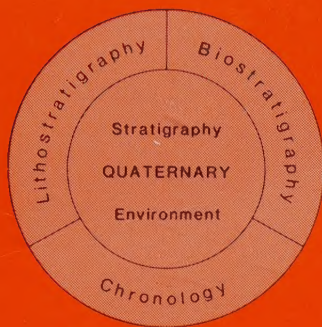
- Aario, R., 1972a: Exposed bed forms and inferred three-dimensional flow geometry in an esker delta, Finland. - 24th Int. Geol. Congr., Section 12, Montreal 1972.
- Aario, R., 1972b: Associations of bed forms and paleocurrent patterns in an esker delta, Haapajärvi, Finland. - Ann. Acad. Sci. Fennicae Ser. A III, Pap. 110, 55 pp.
- Adam, P.P., 1975: Ice ages and the thermal equilibrium of the earth II. - Quat. Res. 5, 161-171.
- Adrielsson, L., Mohrén, E. & Daniel, E., 1981: Beskrivning till jordartskartan Helsingborg SV. - Sver. Geol. Unders. Ae 16, 104 p.
- Agterberg, F.P. & Banerjee, J., 1969: Stochastic model for the deposition of varves in glacial lake Barlow, Ojibway, Ontario, Canada. - Can. J. Earth. Sci. 6, 625-652.
- Allen, J.R.L., 1965: Sedimentation in the lee of small underwater sand waves. - J. Geol. 73, 95-116.
- Allen, J.R.L., 1968: Current ripples and their relation to patterns of water and sediment motion. - North Holland, Amsterdam, 433 pp.
- Allen, J.R.L., 1970: Studies in fluvial sedimentation: A comparison of fining upwards cyclothem, with special reference to coarse-member composition and interpretation. - J. Sediment. Petr. 40, 298-323.
- Allen, J.R.L., 1973: A classification of climbing-ripple cross-lamination. - J. Geol. Soc. London 129, 537-541.
- Andrews, J.T., 1982: Comment on "New evidence from beneath the western North Atlantic for depth of glacial erosion in Greenland and North America" by E.P. Laine. - Quat. Res. 19, 18-37.
- Andrews, J.T. & Mahaffy, M.A., 1975: Rate of growth of the Laurentide ice sheet based on a three-dimensional numerical ice flow model. - In: Abstracts with programs 7, Geol. Soc. Am., 977-978.
- Andrews, J. T. & Miller, G. H., 1979: Glacial erosion and the ice sheet divides, northeastern Laurentide Ice Sheet, on the basis of the distribution of limestone erratics. Geology 7, 592-596.
- Ashley, G.M., 1972: Rhythmic sedimentation in glacial lake Hitchcock, Massachusetts-Connecticut. - Univ. Mass., Amherst, Geology Publ. 10, 148 p.
- Ashley, G.M., 1975: Rhythmic sedimentation in glacial lake Hitchcock, Massachusetts-Connecticut. In: Jopling, A.V. & McDonald, B.C. (eds.): Glaciofluvial and glaciolacustrine sedimentation. - Soc. Econ. Palaeontologists Mineralogists Spec. Publ. 23, 304-320.
- Axelsson, V., 1967: The Laitaure delta. A study of deltaic morphology and processes. - Geogr. Ann. 49A, 3-127.
- Bahnson, H., 1973: Lithological investigations in some Danish boulder-clay profiles. - Bull. Geol. Inst. Univ. Uppsala, New series 5, 93-109.
- Bahnson, H., Petersen, K.S., Konradi, P. & Knudson, K.L., 1974: Stratigraphy of Quaternary deposits in the Skaerumhede II boring: Lithology, molluscs and foraminifera. - Danm. Geol. Unders., Årbog 1973, 27-62.
- Bagnold, R.A., 1946: Motion of waves in shallow water. Interaction between waves and sand bottoms. - Prov. Roy. Soc. London, Ser. A187, 1-18.
- Banham, P.H., 1975: Glaciotectionic structures: a general discussion with particular reference to the contorted drift of Norfolk. In: Wright, A.E. & Moseley, F. (eds.): Ice Ages: Ancient and Modern. - Seal House Press, Liverpool, 69-94.
- Banham, P. H., 1977: Glaciotectionics in till stratigraphy. - Boreas 6, 101-105.
- Berglund, B.E. & Lagerlund, E., 1981: Eemian and Weichselian stratigraphy in South Sweden. - Boreas 8, 323-362.
- Bergström, J. & Shaikh, N. A., 1980: Malmer, industriella mineral och bergarter i Kristianstads län. - Sver. Geol. Unders., Rapp. Medd. 22, 88p.
- Bergström, J. & Shaikh, N. A., 1982: Malmer, industriella mineral och bergarter i Malmöhus län. - Sver. Geol. Unders., Rapp. Medd. 31, 88p.
- Bergström, J., Holland, B., Larsson, K., Norling, E. & Sivhed, U., 1982: Guide to excursions in Scania. - Sver. Geol. Unders. Ca 54, 95 p.
- Birks, H.J.B., 1974: Numerical zonations of Flandrian pollen data. - New Phytol. 73, 351-358.
- Birks, H.J.B. & Berglund, B.E., 1979: Holocene pollen stratigraphy of southern Sweden: a reappraisal using numerical methods. - Boreas 8, 257-279.
- Björck, S., 1979: Late Weichselian stratigraphy of Blekinge, SE Sweden, and water level changes in the Baltic Ice Lake. - Univ. Lund, Dept. Quat. Geol., Thesis 7, 248 p.
- Boersma, J.R., 1967: Remarkable types of mega cross-stratification in the fluvial sequence of a subrecent distributary of the Rhine. Amerongen; The Netherlands. - Geol. Mijnbouw 46, 217-235.
- Boersma, J.R., 1970: Distinguishing features of wave-ripple cross-stratification and morphology. - Doctoral thesis, Univ. Utrecht. 65 p.
- Boersma, J.R., Meene, E.A. van de & Tjalsma, R.C., 1968: Intricate cross-stratification due to interaction of a mega ripple with its lee-side system of backflow ripples (upper-pointbar deposits, lower Rhine). - Sedimentology II, 147-162.
- Boothroyd, J.C. & Ashley, G.M., 1975: Processes, bar morphology and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska. In: Jopling, A.V. & McDonald, B.C. (eds.): Glaciofluvial and glaciolacustrine sedimentation. - Soc. Econ. Palaeontologists Mineralogists Spec. Publ. 23, 193-222.
- Boulton, G.S., 1970a: On the origin and transport of englacial debris in Svalbard glaciers. - J. Geol. 9, 213-229.
- Boulton, G.S., 1970b: On the deposition of subglacial and melt-out tills at the margins of certain Svalbard glaciers. - J. Glaciol. 9, 231-245.
- Boulton, G.S., 1972: Modern Arctic glaciers as depositional models for former ice sheets. - J. Geol. Soc. London 128, 361-393.
- Boulton, G.S., 1972: The role of thermal regime in glacial sedimentation. - Inst. Br. Geogr. Spec. Publ. 4, 1-19.
- Boulton, G.S., 1974: Processes and patterns of glacial erosion. In: Coates, D.R. (ed.): Glacial geomorphology. - State Univ. New York, Binghamton, 41-48.
- Boulton, G.S., 1975: Processes and patterns of subglacial sedimentation: a theoretical

- approach. In: Wright, A.E. & Moseley, F. (eds.): *Ice Ages: Ancient and modern*. - Seel House press, Liverpool, 7-42.
- Boulton, G.S., 1982: Subglacial processes and the development of glacial bedforms. In: Davidson-Arnott, R., Nickling, W. & Fahey, B.D. (eds.): *Research in glacial, glaciofluvial and glacio-lacustrine systems*. - Proceedings of the 6th Guelph Symposium on Geomorphology, 1980. Geo Books, Norwich, 318p.
- Boulton, G.S., Dent, D.L. & Morris, E.M., 1974: Subglacial shearing and crushing and the role of water pressures in tills from south-east Iceland. - *Geogr. Ann.* 56A, 133-145.
- Boulton, G.S., Jones, A.S., Clayton, K.M. & Kennings, M.J., 1977: A British ice-sheet model and patterns of glacial erosion and deposition in Britain. In: Shotton, F. W. (ed.): *British Quaternary studies - recent advances*. - Clarendon Press, Oxford, 231-246.
- Boulton, G.S. & Paul, M.A., 1976: The influence of genetic processes on some geotechnical properties of glacial tills. - *Quat. J. Eng. Geol.* 9, 159-194.
- Bouma, A.H., 1962: *Sedimentology of some flysch deposits. A graphic approach to facies interpretation*. - Elsevier Publ. Comp., 168 p.
- Bouma, A.H., 1972: Fossil contourites in Lower Niesenflysch, Switzerland. - *J. Sediment. Petr.* 42, 917-921.
- Bouma, A.H., 1973: Leveed-channel deposits, turbidities and contourites in deeper part of Gulf of Mexico. - *Transactions - Gulf Coast Assoc. Geol. Soc.* 23, 368-376.
- Bridge, J.S., 1978: Origin of horizontal lamination under turbulent boundary layers. - *Sediment. Geol.* 20, 1-16.
- Cant, D.J. & Walther, R.G., 1978: Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada. - *Sedimentology* 25, 625-648.
- Clayton, L. & Moran, S.R., 1974: A glacial process form model. In: Coates, D.R. (ed.): *Glacial geomorphology*. State Univ. New York, Binghamton, 88-119.
- Coleman, J.M. & Cagliano, S.M., 1965: Sedimentary structures: Mississippi River deltaic plain. In: Middleton, G.V. (ed.): *Primary sedimentary structures and their hydrodynamic interpretation*. - Soc. Econ. Paleontologists Mineralogists. Spec. Publ. 12, 133-148.
- Collinson, J.D., 1970: Bedforms of the Tana river, Norway. - *Geogr. Ann.* 52A, 31-56.
- Davis, J. C., 1973: Statistics and data analysis in geology with Fortran programs Robert J. Sampson. - John Wiley & Sons, New York, 550p.
- De Geer, G., 1912: A geochronology of the last 12 000 years. - *Compte Rendu du XI:e Congres Geologique International*, 241-253.
- De Geer, G., 1940: *Geochronologica Suecia Principes*. - Kungl. Sv. Vetenskapsak. Handl., Ser. 3, band 18, no. 6, 360 p.
- Douglas, D.J., 1962: The structure of sedimentary deposits of braided rivers. - *Sedimentology* 1, 167-190.
- Ehlers, J., 1979: Fine gravel analysis after the Dutch method as tested out on Ristinge Klint, Denmark. - *Bull. geol. Soc. Den.* 27, 157-165.
- Ehlers, J., 1983: Glacial deposits in North-West Europe. - Balkema, Rotterdam, 470 p.
- Erdmann, E., 1873: *Iakttagelser öfver moränbildningar och deraf betäckta skiktade jordarter i Skåne*. - *Geol. Fören. Stockh. Förh.* 2, 13-24.
- Erdmann, E., 1874: *Bidrag till kännedomen om de lösa jordaflageringarne i Skåne II*. - *Geol. Fören. Stockh. Förh.* 2, 101-117.
- Erdmann, E., 1881: *Beskrifning till kartbladet Helsingborg*. - *Sver. Geol. Unders.* Aa 74, 160 p.
- Erdmann, E., 1883: *Bidrag till kännedomen om de lösa jordaflageringarne i Skåne: III. Några profiler på ön Hven och närliggande Skånska kust*. - *Geol. Fören. Stockh. Förh.* 6, 425-434.
- Eyles, N., 1983: Modern Icelandic glaciers as depositional models for 'hummocky moraine' in the Scottish Highlands. In: Evenson, E. B. & Schlüchter, C.H. (eds.): *Tills and related sediments*. - A.A. Balkema, in press.
- Eyles, N., Eyles, C.H. & Miall, A.D., 1983: Lithofacies types and vertical profile models: an alternative approach to the description of environmental interpretation of glacial diamict and diamictite sequences. - *Sedimentology* 30, 393-410.
- Eyles, N. & Menzies, J., 1983: The subglacial landsystem. In: Eyles, N. (ed.): *Glacial geology. An introduction for engineers and earth scientists*. - Pergamon Press, Oxford, 19-70.
- Fisher, R.V., 1971: Features of coarse-grained, high concentration fluids and their deposits. - *J. Sediment. Petrol.* 41, 916-927.
- Flint, R.F., 1957: *Glacial and Pleistocene geology*. - John Wiley and Sons, New York, 553 p.
- Flint, R.F., 1971: *Glacial and Quaternary geology*. - John Wiley and Sons, New York, 892 p.
- Folk, R.L. & Ward, W., 1957: Brazos River bar: A study in the significance of grain size parameters. - *J. Sediment. Petrol.* 27, 3-26.
- Friedman, G.M. & Sanders, J.E., 1978: *Principles of sedimentology*. - John Wiley and Sons, New York, 792 p.
- Gandahl, R., 1952: *Bestämning av kornstorlek med hydrometer*. - *Geol. Fören. Stockh. Förh.* 74, 497-512.
- Gilbert, G. K., 1899: Ripple marks and cross-bedding. - *Geol. Soc. Am. Bull.* 10, 135-140.
- Glen, J.W., 1955: The creep of polycrystalline ice. - *Proc. R. Soc. A228*, 519-538.
- Goldthwait, R. P., 1975: *Glacial deposits. - Benchmark papers in geology* 21, Dowden Hutchinson & Ross Inc., Stroudsburg, United States, 464p.
- Gustavson, T.C., 1975: Sedimentation and physical limnology in proglacial Malaspina lake southeastern Alaska. In: Jopling, A.V. & McDonald, B.C. (eds.): *Glaciofluvial and glaciolacustrine sedimentation*. - Soc. Econ. Paleontologists Mineralogists Spec. Publ. 23, 249-263.
- Gustavson, T.C., Ashley, G.M. & Boothroyd, J.C., 1975: Depositional sequences in glaciolacustrine deltas. In: Jopling, A. V. & McDonald, V. (eds.): *Glaciofluvial and glaciolacustrine sedimentation*. - Soc. Econ. Paleontologists Mineralogists Spec. Publ. 23, 254-280.
- Haldorsen, S. & Shaw, J., 1982: The problem of recognizing melt-out till. - *Boreas* 11, 261-277.
- Hansen, S., 1933: *De glaciala aarsvarv i Skåne*. - *Geol. Fören. Stockh. Förh.* 55, 632-642.
- Hansen, S., 1940: *Varvighed i danske og skaanske senglaciale aflejringer*. - *Dan. Geol. Unders.* 11, 63, 478 p.
- Hedberg, H.D., 1976: *International stratigra-*

- phic guide - A guide to stratigraphic classification, terminology and procedure. - John Wiley and Sons, New York, 200 p.
- Hein, F.J. & Walker, R.G., 1977: Bar evolution and development of stratification in the gravelly, braided Kicking Horse River, British Columbia. - *Can. J. Earth. Sci.* 14, 562-570.
- Hesse, R. & Chough, S.K., 1980: The Northwest Atlantic mid-ocean channel of the Labrador Sea: 11. Deposition of parallel laminated levee muds from the viscous sublayer of low density turbidity currents. - *Sedimentology* 27, 697-711.
- Hills, E. S., 1972: Elements of structural geology. Second edition. - John Wiley & Sons, New York, 502p.
- Hjulström, F., 1939: Transportation of detritus by moving water. In: Trask, P. D. (ed.): Recent marine sediments. - Am. Assoc. Petrol. Geol., 5-31.
- Holmström, L.P., 1865: Iakttagelser öfver märken i Skåne efter istiden. - *Chronholmska Borttryckeriet, Malmö*, 25 p.
- Holmström, L.P., 1874: Bidrag till kännedomen af moränbildningarna på Hven och närliggande Skånska Kust. - *Geol. Fören. Stockh. Förh.* 2, 96-101.
- Holmström, L.P., 1904: Öfversikt af den glaciala afslipningen i sydsandinavien. - *Geol. Fören. Stockh. Förh.* 26, 241-432.
- Holst, N.O., 1911: Alnarps-floden, en svensk Cromer-flod. - *Sver. Geol. Unders.* C237, 64 p.
- Houmark-Nielsen, M., 1980: Glacialstratigrafi en omkring det nordlige Baelthav. - Unpubl. lic. thesis. Inst. Gen. Geol., Univ. Copenhagen, 190 p.
- Hörnsten, Å., 1979: Sand och övriga jordarter i Öresund. Kommentar till SGU's marin-geologiska karta över Öresund. - *Sver. Geol. Unders., Rapp. Medd.* 13.
- Inman, D.L., 1952: Measures for describing the size distribution of sediments. - *J. Sediment. Petrol.* 22, 125-145.
- Ives, J.D., Andrews, J.T. & Barry, R.G., 1975: Growth and decay of the Laurentide ice sheet and comparisons with Fennoscandia. - *Die Naturwiss.* 62, 118-125.
- Jackson, R.G., 1976: Largescale ripples of the Lower Wabash River. - *Sedimentology* 23, 593-623.
- Johnsson, G., 1956: Glaciomorfologiska studier i södra Sverige. - *Lindstedts, Lund*, 407 p.
- Johnsson, G., 1958: Submoraine ice-wedges in western Scania. - *Geol. Fören. Stockh. Förh.* 80, 333-339.
- Johnsson, G., 1962: Periglacial phenomena in southern Sweden. - *Geogr. Ann.* 44, 378-404.
- Johnsson, G., 1963: Periglacial ice-wedge polygons at Hässleholm, southernmost Sweden. - *Sv. Geogr. Årsb.* 39, 173-176.
- Johnsson, G., 1984: Ice-wedge polygons and other periglacial phenomena in intermoraine sediments at Öresund, South Sweden. - *Geol. Fören. Stockh. Förh.* 105, in press.
- Jopling, A.V., 1961: Origin of regressive ripples explained in terms of fluid-mechanic processes. - *Short Papers in Geol. and Hydrol. Sci., U.S. Geol. Surv. Prof. Paper* 424-D, 15-17.
- Jopling, A. V., 1962: Mechanics of small-scale delta formation. - a laboratory study. - *Nat. Shallow Water Res. Conf. 1st., Proc. Nat. Sci. Foundation, Washington D. C.*, 291-295.
- Jopling, A.V., 1963: Hydraulic studies on the origin of bedding. - *Sedimentology* 2, 115-121.
- Jopling, A.V., 1965: Laboratory study of sorting processes in cross-bedded deposits. In: Middleton, G.V. (ed.): Primary sedimentary structures and their hydrodynamic interpretation. - *Soc. Econ. Paleontologist Mineralogists Spec. Publ.* 12, 53-65.
- Jopling, A.V. & Walker, R.G., 1968: Morphology and origin of ripple-drift cross lamination, with examples from the Pleistocene of Massachusetts. - *J. Sediment. Petrol.* 38, 971-984.
- Krumbein, W. G., 1933: The dispersion of fine-grained sediments for mechanical analysis. - *J. Sediment. Petrol.* 3, 121-135.
- Keunen, Ph. H., 1951a: Turbidity current as the cause of glacial varves. - *J. Geol.* 59, 507-508.
- Keunen, Ph. H., 1951b: Mechanics of varve formation and the action of turbidity currents. - *Geol. Fören. Stockh. Förh.* 73, 69-84.
- Lagerlund, E., 1971: Some aspects on till stratigraphy in the light of lithostratigraphic studies on the Kullen peninsula in northwestern Scania. - *Geol. Fören. Stockh. Förh.* 93, 653-658.
- Lagerlund, E., 1977a: Förutsättningar för moränstratigrafiska undersökningar på Kullen i Nordvästskåne - teoriutveckling och neotektonik. - *Univ. Lund, Dept. Quat. Geol., Thesis* 5, 160 p.
- Lagerlund, E., 1977b: Till studies and neotectonics in northwest Skåne, South Sweden. - *Boreas* 6, 159-166.
- Lagerlund, E., 1980: Proglacial till in western Scania, Sweden - origin and classification problems. In: Sollid, J.L. (ed.): INQUA Till Norway 1979. - *Nor. Geogr. Tidskr.* 34, p. 98.
- Lagerlund, E., 1983: The pleistocene stratigraphy of Skåne, southern Sweden. In: Ehlers, J. (ed.): Glacial deposits in Northwest Europe. - *A.A. Balkema Rotterdam*, 155-159.
- Lawson, D.E., 1979: Sedimentological analysis of the western terminus region of the Matanuska Glacier, Alaska. - *Cold Regions Res. Eng. Lab., Rept* 79-9.
- Lowe, D.R., 1975: Water-escape structures in coarse-grained sediments. - *Sedimentology* 22, 157-204.
- Lowe, D.R. & Lopicollo, R.D., 1974: The characteristics and origins of dish and pillar structures. - *J. Sediment. Petrol.* 44, 484-501.
- Lundqvist, J., 1952: Bergarterna i Dalamörnernas block- och grusmaterial. - *Sver. Geol. Unders.* C 525, 48 p.
- Madsen, V., 1917: En kvartær dislokation ved Sundsvik tegelbruk i Skåne. - *Geol. Fören. Stockh. Förh.* 39, 597-602.
- Mangerud, J., Andersen, S. T., Berglund, B. E. & Donner, J., 1974: Quaternary stratigraphy of Norden, a proposal for terminology and classification. - *Boreas* 3, 109-128.
- Marcussen, I., 1973: Studies of flow till in Denmark. - *Boreas* 2, 213-231.
- Markgren, M., 1961: Glaciale tektoniken i Vens och Glumslövsområdets strandkintar. - *Sv. Geogr. Årsb.* 37, 115-123.
- Mattson, Å., 1960: Die glaziale Felsskulptur und die jünsten Eisströme über Süd-Schonen-Nordbornholm. - *Z. Geomorphologie* 4:2, 129-140.
- May, R.W. & Dreimanis, A., 1976: Differentiation of glacial tills in southern Ontario, Canada, based on their Cu, Zn, Cr, and Ni geochemistry. - *Geol. Soc. Am. Mem.*, No. 136, 221-228.
- McKee, E.D., 1965: Experiments on ripple lamination. In: Middleton, G.V. (ed.): Primary sedimentary structures and

- their hydrodynamic interpretation. - Soc. Econ. Paleontologists Mineralogists Spec. Publ. 12, 66-83.
- Miall, A.D., 1977: A review of the braided river depositional environment. - *Earth Sci. Rev.* 13, 1-62.
- Miall, A.D., 1983: Principles of sedimentary basin analysis. - Springer Verlag, New York.
- Middleton, G.V. & Hampton, M.A., 1973: Sediment gravity flows: mechanisms of flow and deposition. - Soc. Econ. Paleontologists Mineralogists, Pacific Sect., Short Course Turbidities and deep Water Sedimentation, 38 p.
- Middleton, G.V. & Hampton, M.A., 1976: Subaqueous sediment transport and deposition by sediment gravity flow. In: Stanley, D.J. & Swift, D.J.P. (eds.): Marine sediment transport and environmental management. - Wiley - Intersci. Publ., New York, 197-218.
- Miller, U., 1971: Microfossils in the Quaternary layers at Tofthög, southern Scania. - *Geol. Fören. Stockh. Förh.* 93, 507-524.
- Miller, U., 1977: Pleistocene deposits of the Alnarp Valley, southern Sweden. - Microfossils and their stratigraphical application. - Univ. Lund., Dept. Quat. Geol., Thesis 4, 116p.
- Milthers, V., 1922: Nordöstsjaellands geologi. Danm. Geol. Unders., Raekke 5, 3, 182p.
- Moore, D.G., 1961: Submarine slumps. - *J. Sediment. Petrol.* 31, 343-357.
- Moore, D.G., 1969: Reflection profiling studies of the California continental borderland: Structure and quaternary turbidite basins. - *Geol. Soc. Am. Spec. Papers.* 107, 1-142.
- Moran, S.R., Clayton, L., Hooke, R. LeB.-Fenton, M.M. & Andriashek, L.D., 1980: Glacier-bed landforms of the prairie region of the North America. - *J. Glacial.* 25, 457-476.
- Morgenstern, N.R., 1967: Submarine slumping and the initiation of turbidity currents. In: Richards, A.F., (ed.): Marine geotechnique. - Univ. Illinois Press, 189-220.
- Morrison, R.B., 1967: Principles of Quaternary soil stratigraphy. In: Morrison, R.B., & Wright, H.E. (eds.): Quaternary soils. - Int. Ass. Quat. Res. 9th. Cong., Proc. 9. Reno, Nevada, Univ. Nevada, Desert Research Inst., Center for Water Resources Research, 1-69.
- Munthe, H., 1896: Studien über ältere Quartärablagerungen im südbaltischen Gebiet. - *Bull. Geol. Ints. Univ. Uppsala* 5,
- Mörner, N.-A., 1969: The Late Quaternary history of the Kattegatt sea and the Swedish West Coast. - *Sver. Geol. Unders.* C 640, 487 p.
- Nilsson, K., 1971: Skivarpsströmmen. - Grundvattenförekomst i sydvästra Skåne.- Symposium vid Lunds Tekniska Högskola 7-8 juni 1971, 61-67.
- Nilsson, K., 1973: Glacialgeologiska problem i Sydvästraskåne. - Univ. Lund, Dept. Quat. Geol., Thesis 1, 20 p.
- Nilsson, T., 1983: The Pleistocene: Geology and life in the Quaternary ice age. - Reidel/MTP Press, 650 p.
- Nobles, L. H. & Weertman, J., 1971: Influence of irregularities of the bed of an ice sheet on deposition rate of till. In: Goldthwait, R. P., (ed.): Till, a symposium. - Ohio State Univ. Press, 117-126.
- Nye, J. F., 1973: The motion of ice past obstacles. In: Whalley, E., Jones, S. J. & Gold, L. W., (eds.): Physics and chemistry of ice. - R. Soc. Can., Ottawa, 387-395.
- Okko, V., 1955: Glacial drift in Iceland, its origin and morphology. - *Bull. Comm. Géol. Finl.* 170, 1-133.
- Paterson, W.S.B., 1969: The physics of glaciers. - Pergamon Press, Oxford, 250 p.
- Paul, M.A., 1983: The supraglacial landscape. In: Eyles, N. (ed.): Glacial geology. An introduction for engineers and earth scientists. - Pergamon Press, Oxford, 71-90 p.
- Petersen, K.S. & Konradi, P.B., 1974: Lithologiske og palaeontologiske beskrivelser af profiler i kvartaeret på Sjaelland. - Danm. Geol. Foren., Årsskrift 1973, 47-56.
- Péwé, T.L., 1963: Ice wedges in Alaska - classification, distribution and climatic significance. - *Geol. Soc. Am. Spec. Pap.* 76, 129 p.
- Potter, P.E. & Pettijohn, F.J., 1963: Paleocurrents and basin analysis. - Springer Verlag, Berlin, 296 p.
- Price, R.J., 1969: Moraines, sandar, kames and eskers near Breidamerjökull, Iceland. - *Trans. Inst. Brit. Geogr.* 46, 17-43.
- Price, R.J., 1970: Moraines at Fjällsjökull, Iceland. - *Arctic and Alpine Res.* 2, 27-42.
- Rahn, P.H., 1967: Sheetfloods, streamfloods, and the formation of pediments. - *Ass. Am. Geogr. Ann.* 57, 593-604.
- Rasmussen, L.Aa., 1973a: The Quaternary stratigraphy and dislocations on Ven. - *Bull. Geol. Inst. Univ. Uppsala*, N.S. 5, 37-40.
- Rasmussen, L.Aa., 1973b: The Quaternary stratigraphy and dislocations on Ven. - *Bull. Geol. Inst. Univ. Uppsala*, N.S. 5, 37-39.
- Rasmussen, L. Aa., 1973c: Glacialgeologien på Hven og den nærliggende skånske kyst. - Institut for almen Geologi, Copenhagen, Unpublished thesis, 136p.
- Rasmussen, L.Aa., 1974: Om morænestratigrafi i det nordlige Öresundområde. - Danm. Geol. Foren. Årsskrift 1973, 132-139.
- Rose, J., 1974: Small scale variability of some sedimentary properties of lodgement and slumped till. - *Proc. Geol. Ass.* 85, 223-237.
- Schlüchter, C., (ed.), 1982: Report on activities 1977-1982, INQA Commission on Genesis and Lithology of Quaternary Deposits. - ETH - Zürich, July 1982.
- Shaw, J., 1977a: Tills deposited in arid, polar environments. - *Can. J. Earth. Sci.* 14, 1239-1245.
- Shaw, J., 1977b: Sedimentation in an alpine lake during deglaciation, Okanagan Valley, British Columbia, Canada. - *Geogr. Ann.* 59A, 221-240.
- Simons, D.B. & Richardson, E.V., 1961: Forms of bed roughness in alluvial channels. - *Am. Soc. Civ. Engrs.*, Proc. HY3 87, 87-105.
- Simons, D.B. & Richardson, E.V., 1962: Resistance to flow in alluvial channels. - *Am. Soc. Civ. Engrs.*, Trans. 127, 927-953.
- Simons, D.B., Richardson, E.V. & Nordin, C. F., 1965: Sedimentary structures generated by flow in alluvial channels. In: Middleton, G.V. (ed.): Primary sedimentary structures and their hydrodynamic interpretation. - Soc. Econ. Paleontologists Mineralogists, Spec. Publ. 12, 34-52.
- Smith, N.D., 1970: The braided stream depositional environment: comparison of the Platte River with some Silurian clastic rocks, north-central Appalachians. - *Geol. Surv. Am. Bull.* 81, 2993-3014.
- Smith, N.D., 1971: Transverse bars and brai-

- ding in the lower Platte River, Nebraska. - Geol. Surv. Am. Bull. 82, 3407-3420.
- Smith, N.D., 1972: Some sedimentological aspects of planar cross-stratification in a sandy braided river. - J. Sedimentary Petrology 42, 624-634.
- Sorgenfrei, Th., 1945: Traek af Alnarpsdalens geologiske opbygning. - Danm. Geol. Foren., 617 p.
- Sorgenfrei, Th., 1971: En sammenfattende oversigt over de hydrologiske forhold i SV-Skåne og Sjaelland. - Grundvattenforekomster i sydvästra Skåne, Symposium vid Lunds Tekniska Högskola 7-8 juni 1971, 23-35.
- Southard, J.B., 1975: Bed configuration. - Soc. Econ. Paleontologists Mineralogists Short Course 2, 5-44.
- Stanley, K. O., 1974: Morphology and hydraulic significance of climbing ripples with superimposed micro-ripple-drift cross-lamination in Lower Quaternary lake silts, Nebraska. - J. Sediment. Petrol. 44, 427-483.
- Stephan, H.-J. & Ehlers, J., 1979: Forms at the base of till strata as indicators of ice movement. - J. Glaciol. 87, 345-355.
- Stow, D. A. W. & Bowen, A. J., 1978: Origin of lamination in deep-sea fine-grained sediments. - Nature 274, 324-328.
- Strömqvist, L., 1973: Geomorfologiska studier av blockhav och blockfält i norra Skandinavien. - Uppsala Univ. Naturgeogr. Inst. Rept. 22, 161p.
- Sugden, D. E. & John, B. S., 1976: Glaciers and landscape. - Edward Arnold Ltd., London, 376p.
- Sundborg, Å., 1956: The river Klarälven: A study of fluvial processes. - Geogr. Ann. 38, 125-316.
- Sundborg, Å., 1967: Some aspects on fluvial sediments and fluvial morphology. I. General views and graphic methods. - Geogr. Annlr. 49, 333-342.
- Terzagi, K. & Peck, R. B., 1967: Soil mechanics in engineering practice. - John Wiley & Sons, New York, 729p.
- Torell, O., 1864: Unpublished manuscript reproduced in Torell, O., 1873: Undersökningar öfver istiden del II, 51-62.
- Torell, O., 1872: Undersökningar öfver istiden del I. - Aftryck ur Öfversigt af K. Vetenskapsakad. Förhandl. 1872, P. A. Nordstedt & Söner, Stockholm, 44p.
- Torell, O., 1873: Undersökningar öfver istiden del II. Skandinaviska landisens uträkning under isperioden. - Öfversigt af K. Vetenskapsakad. Förhandl. 1873, no. 1, 47-64.
- Tornquist, A., 1908: Die Feststellung des Südwestrandes des baltisch-russischen Schildes und die geoteknische Zugehörigkeit der ost-preussischen Scholle. - Schriften der Physikalisch-Ökonomischen Gesellschaft 49 (1), 1-12, Königsberg.
- Vivian, R., 1970: Hydrologie et erosion sous-glaciers. - Revue Géogr. alp. 58, 241-264.
- Walcott, R.I., 1970: Isostatic response to loading of the crust in Canada. - Can. J. Earth Sci. 7.
- Walker, R. G., 1963: Distinctive types of ripple-drift cross-lamination. - Sedimentology 2, 173-188.
- Walker, R. G., 1975: Generalized facies models for resedimented conglomerates of tubidite association. - Geol. Soc. Am. Bull. 86, 737-748.
- Weertman, J., 1957: On the sliding of glaciers. J. Glaciol. 3, 33-38.
- Weertmann, J., 1964: The theory of glacier sliding. - J. Glaciol. 5, 287-303.
- Wennberg, G., 1949: Differentialrörelser i landisen. - Carl Bloms boktryckeri AB, Lund, 201p.
- Westgate, J. A., 1968: Linear sole markings in Pleistocene till. - Geol. Mag. 105, 501-505.
- Wright, L. D., 1977: Sediment transport and deposition at river mouths: a synthesis. - Geol. Soc. Am. Bull. 88, 837-868.
- Zandstra, J. G., 1978: Einführung in die Feinkiesanalyse. - Der Geschiebesammler 12, 21-38.
- Ödum, H., 1933: Marint interglacial på Sjaelland, Hven, Mön och Rugen. - Dan. Geol. Unders. 4, 2, 44p.
- Örstedt, A. S., 1844: De regionibus marinis. - Hauniae MDCCCXLV.



LUNDQUA Thesis

At the Department of Quaternary Geology, Lund University, three series are published, named "Thesis", "Report" and "Uppdrag". The "Thesis" series contains doctor dissertations, the "Report" series primary material, often of a monographic character, which cannot be published in extenso in ordinary scientific journals. The "Uppdrag" series contains selected examples of expert reports, generally in Swedish, which may be of some general interest.

The "Thesis" and the "Report" series cover different aspects of Quaternary stratigraphy and environment - methods, lithostratigraphy, biostratigraphy, chronology, as well as their applications within technical geology, resource geology, nature conservancy, archaeology etc. Complete lists of publications may be ordered from the Dept. of Quaternary Geology, Tornavägen 13, S-223 63 Lund, Sweden.

Editorial committee: Björn E. Berglund, Svante Björck, Gunnar Digerfeldt, Christian Hjort and Erik Lagerlund.
Technical editor: Per Sandgren.

1. Nilsson, K. 1973: Glacialgeologiska problem i Sydvästska (English summary: Problems of glacial geology in southwestern Scania). 5 S.kr.
2. Bjelm, L. 1976: Deglaciationen av Småländska höglandet, speciellt med avseende på deglaciationsdynamik, ismaktighet och tidsställning (Deglaciation of the Småland Highland, with special reference to deglaciation dynamics, icethickness and chronology). 15 S.kr.
3. Göransson, H. 1977: The Flandrian Vegetational History of Southern Östergötland. 15 S.kr.
4. Miller, U. 1977: Pleistocene deposits of the Alnarp Valley, southern Sweden. - Microfossils and their stratigraphic application. 15 S.kr.
5. Lagerlund, E. 1977: Förutsättningar för moränstratigrafiska undersökningar på Kullen i Nordvästskåne - teoriutveckling och neotektonik (Till-stratigraphical studies in the Kullen area, NW Skåne, South Sweden: basic theory and neotectonics. 15 S.kr.
6. Hildén, A. 1979: Deglaciationsförloppet i trakten av Berghems moränen, öster om Göteborg (English summary: The deglaciation in the vicinity of the Berghem Moraine east of Göteborg). 20 S.kr.
7. Björck, S. 1979: Late Weichselian stratigraphy of Blekinge, SE Sweden, and water level changes in the Baltic Ice Lake. 20 S.kr.
8. Andersson, O.H. 1981: Borrning och dokumentation. Borrningsteknik jämte metodik för geologisk datainsamling under borrhings gång (Drilling and documentation). 30 S.kr.
9. Hjort, Ch. 1981: Studies of the Quaternary in Northeast Greenland. 20 S.kr.
10. Hjelmroos-Ericsson, M. 1981: Holocene development of Lake Wielkie Gacho area, north-western Poland. 30 S.kr.
11. Liljegren, R. 1982: Paleoeкологи och strandförskjutning i en Littorinavik vid Spjälkö i mellersta Blekinge (English summary: Palaeoecology and shore displacement in a Littorina bay at Spjälkö, Blekinge). 30 S.kr.
12. Norddahl, H. 1983: Late Quaternary stratigraphy of Fnjóskadalur central North Iceland, a study of sediments, ice-lake strandlines, glacial isostasy and ice-free areas. 30 S.kr.
13. Robison, J.M. 1983: Glaciofluvial sedimentation: A key to the deglaciation of the Laholm area, southern Sweden. 35 S.kr.
14. Sandgren, P. 1983: The deglaciation of the Klippan area, southern Sweden, a study of glaciofluvial and glaciomarine sediments. 35 S.kr.
15. Åmark, M. 1984: The deglaciation of the eastern part of Skåne, southern Sweden. A study of till and stratified drift. 35 S.kr.

CODEN: LUNBDS/(NBGK-1016)/1-126/(1984)

ISSN: 0346-8976

Printed by Tid-Tryck Västergatan, Malmö